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CONTENTS

ARTICLES

(Figures in bold type denote number of issue)

ADAMS, C. G.
A Note on the Age of the Laig Gorge Beds, Eigg, **4**, 322–325.

AGER, D. V.
Nomenclatural Problems in the Mesozoic Rhynchonelloidea, **2**, 157–162.

ALLEN, ADRIAN.
Seismic Refraction Investigations of the Preglacial Valley of the River Teifi near Cardigan, **4**, 276–282.

ALLEN, J. R. L.
Cornstone, **1**, 43–48.

ANDERSON, J. G. C.
The Wenlock Strata of South Mayo, **4**, 265–275.

ANDERSON, T. B. D., E. B. BATES, and R. S. HAMILTON.
A Note on the Disposition of Fold Axes in the Silurian Strata around the Newry Granodiorite, **3**, 239–244.

BADEN-POWELL, D. F. W.
On the Nature of the Coralline Crag, **2**, 123–132.

BERRY, F. G.
A New Instrument for Determining Planar Orientation **4**, 309–312.

BROWN, P. E.
Structural Behaviour of Quartzite in a Migmatite Area, Tanganyika, **3**, 250–254.

CAVE, R.
A New Species of *Tretaspis* from South Wales, **4**, 334–337.

CIFELLI, RICHARD.
Notes on the Distribution of English Bathonian Foraminifera, **1**, 33–42.

COORAY, P. G., and P. S. KUMARAPELI.
Corundum in Biotite-Sillimanite Gneiss from near Polgahawela, Ceylon, **6**, 480–487.

CRAMPTON, C. B.
Petrography of the Mesozoic Succession of South Wales, **3**, 215–228.

CURRY, DENNIS.
Eocene Limestones to the West of Jersey, **4**, 289–298.

DEAN, W. T.
The Ordovician Rocks of the Chatwall District, Shropshire, **2**, 163–171.

DINELEY, D. L.
The Old Red Sandstone of Eastern Ekmanfjorden, Vesrspitsbergen, **1**, 18–32.

ELWELL, R. W. D., R. R. SKELHORN, and A. R. DRYSDALL.
Inclined Granitic Pipes in the Diorites of Guernsey, **2**, 89–105.

FARRAND, M. G.
The Distribution of some Elements across the Contacts of Four Xenoliths, **6**, 488–493.

GOBBETT, D. J. A.
A New Species of Trilobite from the Lower Oslobreen Limestone, **6**, 457–460.

GOBBETT, D. J., and C. B. WILSON.
The Oslobreen Series, Upper Hecla Hoek of Ny Friesland, Spitsbergen, **6**, 441–457.

HAMILTON, E.
The Distribution of Radioactivity in some Fine-grained Igneous Rocks, **3**, 255–260.

HOBBS, D. W.
The Strength and Stress-strain Characteristics of Oakdale Coal under Triaxial Compression, **5**, 422–435.

HUDSON, J. D.
The Laig Gorge Beds, Isle of Eigg, **4**, 313–321.
HUDSON, R. G. S.
The Permian and Trias of the Oman Peninsula, Arabia, **4**, 299–308.
HUGHES, CHARLES, J.
An Occurrence of Tilleyite-bearing Limestones in the Isle of Rhum, Inner Hebrides, **5**, 384–388.
JAEGAR, J. C.
Shear Failure of Anisotropic Rocks, **1**, 65–72.
JENKINS, GLYNDWR.
Deformation Features in Cataclastic Quartz, **1**, 73–81.
JOPLIN, GERMAINE, A.
On the Tectonic Environment of Basic Magma, **5**, 363–368.
KING, BASIL, C.
The Form of the Beinn An Dubhaich Granite, Skye, **4**, 326–333.
LJUNGGREN, PONTUS.
A Formation of Marekanite at El Fiscal, Guatemala, **1**, 49–52.
MIDDLETON, GERARD, V.
Spilitic Rocks in South-east Devonshire, **3**, 192–207.
NAHA, K., and S. K. GHOSH.
Archaean Palaeogeography of Eastern and Northern Singhbhum, Eastern India, **5**, 436–439.
NEALE, J. W.
The Subdivision of the Upper D Beds of the Speeton Clay of Speeton, East Yorkshire, **5**, 353–362.
NORMAN, TEOMAN, N.
Azimuths of Primary Linear Structures in Folded Strata, **4**, 338–343.
O'HARA, M. J.
A Garnet Hornblende-Pyroxene Rock from Glenelg, Inverness-shire, **2**, 145–156.
Co-existing Pyroxenes in Metamorphic Rocks, **6**, 498–503.
PITCHER, W. S., and H. H. READ.
Early Transverse Dykes in the Main Donegal Granite, **1**, 53–61.
RAY, P. S.
Ignimbrite in the Kilchrist Vent, Skye, **3**, 229–238.
RHODES, F. H. T.
The Efficacy of Natural Selection as an Evolutionary Agent, with Particular Reference to the Fossil Record, **3**, 177–191.
ROLFE, W. D. IAN.
A Fine Air-Heave Structure from the Old Red Sandstone of Lanarkshire, Scotland, **2**, 133–136.
RUDWICK, M. J. S.
The Feeding Mechanisms of Spire-bearing Fossil Brachiopods, **5**, 369–383
SANDERS, JOHN, E.
Origin of Convoluted Laminae, **5**, 409–421.
SARJEANT, W. A. S.
New Hystrichospheres from the Upper Jurassic of Dorset, **2**, 137–144.
SHOTTON, F. W.
Large Scale Patterned Ground in the Valley of the Worcestershire Avon, **5**, 404–408.
SMITH, SHELAGH M.
Metamorphism of the Jurassic Rocks of Glas Bheinn Bheag, near Dunan, Isle of Skye, **6**, 466–479.
SOPER, N. J.
A Note on Current-bedded Moines near Garve, Wester Ross, **6**, 504–508.
SUTTON, JOHN, and JANET WATSON.
Sedimentary Structures in the Epidotic Grits of Skye **2**, 106–122.
TILLEY, C. E.
Kilauea Magma, 1959–60, **6**, 494–497.

TOCHER, F. E.
Hornblende orientation, **6**, 461–465.
TUCKER, E. V.
Ludlovian Biotite-Bearing Bands, **3**, 245–249.
WALKER, G. P. L.
An Occurrence of Mugearite in Antrim, **1**, 62–64.
WEIR, J. A.
Mudstone Inclusions in Salopian Conglomerates from County Clare **4**, 283–288.
WELLS, ALAN J.
Cyclic Sedimentation : A Review, **5**, 389–403.
WILLIAMS, EMYR.
Intra-Stratal Flow and Convolute Folding, **3**, 208–214.
WILSON, ALLAN F.
Co-Existing Pyroxenes : Some Causes of Variation and Anomalies in the Optically Derived Compositional Tie-Lines, with Particular Reference to Charnockitic Rocks, **1**, 1–17.

REVIEWS

The Earth and its Gravity Field, **1**, 86.
Elements of Geology, **1**, 86.
Principles of Geology, **1**, 86.
Darwin's Biological Work : Some Aspects Reconsidered, **1**, 87.
Introduction to Historical Geology, **2**, 176.
The Pleistocene of Ceylon, **2**, 176.
Beaches and Coasts, **3**, 264.
Mineralogy—Concepts, Descriptions, Determinations, **4**, 349.
Geologic Evolution of Europe, **4**, 350.
British Caenozoic Fossils (Tertiary and Quaternary), **4**, 350.
Basic Geology for Science and Engineering, **5**, 440.
Stratigraphic Principles and Practice, **5**, 440.
The Evolution of North America, **6**, 525.
The Earth Beneath the Sea, **6**, 526.
Structural Methods for the Exploration Geologist, **6**, 527.
Subsurface Mapping, **6**, 527.
Elements of Cartography, **6**, 527.
New Instruments and Methods of Engineering Geology, **6**, 528.
Geology for Engineers, **6**, 528.
Publications Received, **1**, 87–88 ; **4**, 351–352 ; **6**, 529–532.

CORRESPONDENCE

Ultrabasic Pillow Lavas from Cyprus, **1**, 82.
Carboniferous Limestone-Namurian Junction in Ireland, **1**, 84.
The Bugle Pit, Hartwell, near Aylesbury, Bucks, SP(42)793121, **1**, 84.
The Southern Upland Fault in Ireland, **2**, 172.
Cenomanian Ammonite Zones, **2**, 173.
Devonian-Carboniferous Boundary, **2**, 175.
Impregnation of Rocks for Sectioning, **3**, 261.
Festoon Bedding and " Mud-with-Lenticles " Lithology, **3**, 261.
The Old Red Sandstone of Eastern Ekmanfjorden, **3**, 263.
Metamorphic Segregation Bodies, **4**, 344.
Irish Erosion Surfaces, **4**, 345.
Petrography of the Mesozoic Succession of South Wales, **4**, 348.
Deposition of the Old Red Sandstone, **6**, 509.
Wenlock Strata of South Mayo, **6**, 510.
Depth to Top of Postulated Weardale Granite, **6**, 511.
Feeding Mechanisms of Spire-bearing Brachiopods, **6**, 514.
On *Gryphaea*, **6**, 518.
" Deltaic " Cyclothems, **6**, 524.

PLATES

I. Marekanite from Guatemala. **1,** *facing* p. 52.

II. Cataclastic Quartz. **1,** *facing* p. 74.

III. Cataclastic Quartz. **1,** *facing* p. 76.

IV. Inclined Granitic Pipes. **2,** *facing* p. 104.

V. Air-Heave Structure in Old Red Sandstone. **2,** *facing* p. 136.

VI. Hystrichospheres from the Upper Jurassic. **2,** *facing* p. 144.

VII. Ignimbrite in the Kilchrist Vent, Skye. **3,** *facing* p. 238.

VIII. Gneiss and Quartzite, E. of Igawa, Tanganyika. **3,** *facing* p. 254.

IX. Trias of Oman, Arabia. **4,** *facing* p. 308.

X. A New *Tretaspis* from S. Wales. **4,** *facing* p. 336.

XI. Large-scale Patterned Ground. **5,** *facing* p. 408.

XII. Fractured Specimens of Oakdale Coal. **5,** *facing* p. 435.

XIII, XIV. Oslobreen Series. **6,** *facing* p. 460.

XV. *Hystricurus wilsoni* sp. nov. **6,** *facing* p. 460.

XVI. Metamorphism of Jurassic Rocks, Isle of Skye, **6,** *facing* p. 479.

XVII. Current-bedded Moine Psammite, near Garve, Wester Ross. **6,** *facing* p. 508.

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CONTENTS

PAGE

WILSON, ALLAN F. Co-Existing Pyroxenes: Some Causes of Variation and Anomalies in the Optically Derived Compositional Tie-Lines, with Particular Reference to Charnockitic Rocks 1

DINELEY, D. L. The Old Red Sandstone of Eastern Ekmanfjorden, Vestspitsbergen 18

CIFELLI, RICHARD. Notes on the Distribution of English Bathonian Foraminifera 33

ALLEN, J. R. L. Cornstone 43

LJUNGGREN, PONTUS. A Formation of Marekanite at El Fiscal, Guatemala 49

PITCHER, W. S., and READ, H. H. Early Transverse Dykes in the Main Donegal Granite 53

WALKER, G. P. L. An Occurrence of Mugearite in Antrim . . 62

JAEGER, J. C. Shear Failure of Anistropic Rocks 65

JENKINS, GLYNDWR. Deformation Features in Cataclastic Quartz . 73

CORRESPONDENCE
pp. 82–85

REVIEWS
pp. 86–87

PUBLICATIONS RECEIVED
pp. 87–88

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Co-Existing Pyroxenes: Some Causes of Variation and Anomalies in the optically Derived Compositional Tie-Lines, with Particular Reference to Charnockitic Rocks

By ALLAN F. WILSON

ABSTRACT

Available data from charnockitic rocks (including twenty-eight new pyroxene pairs from Australia) suggest that a major cause of variation in optically derived compositional tie-line trends of co-existing pyroxenes is related to the geothermal history of the rocks concerned. It may yet be possible to subdivide the granulite facies, and also to detect mineral instability by critical study of tie-line trends. Recommendations are made whereby mineralogic and structural studies of charnockitic rocks can be integrated to yield more reliable chemical and optical data.

INTRODUCTION

CONSIDERABLE interest has been displayed in the observation of Hess (1941, p. 585) that tie-lines of co-existing pyroxenes from mafic magmas tend to intersect the En–Wo composition line at about En25 Wo75 in the system Wo-En-Fs. Five or six years ago, I noticed that the optical properties of co-existing orthopyroxenes and clinopyroxenes from Central Australian charnockitic granulites indicated an intersection of the Wo–En line consistently much different from En25 Wo75, whereas the pyroxene-bearing igneous rocks of the area, which clearly injected the basement granulites, showed an intersection consistently close to En25 Wo75.

It was thought at the time that this striking feature would eventually prove of value as a means of estimating the difference in temperature of formation of these two groups of rocks, and would also help to distinguish between magmatic and metamorphic rocks in deep-seated zones. It was with some concern, therefore, that I read that Muir and Tilley (1958, p. 406) have concluded that "in metamorphic assemblages carrying two pyroxenes the tie-line trends [chemically derived] show no significant departure from that recorded from plutonic igneous assemblages". Examination of their valuable data (which included five new chemical analyses of co-existing pyroxene pairs), and a survey of all other available data suggest to me that the evidence is as yet insufficient to come to this conclusion.

Table 1 displays all data (known to me) on chemically analysed metamorphic pyroxene pairs where the optical data are reasonably

self-consistent, that is, where 2V determined directly agrees with 2V calculated from refractive indices. The tie-line trends chemically derived of five other pyroxene pairs have been studied, together with those of the nine pyroxene pairs of Table 1. These five pairs show optical data of insufficient self-consistency (as defined above). Table 2 displays other metamorphic pyroxene pairs for which there is sufficient optical data to estimate composition. For this paper the curves of Hess (1949, Pl. 1 and 2; 1952, Fig. 2) were used. Table 3 is a legend for Tables 1, 2, 4, and 5, and includes some comments pertinent to the present study. Some of the data from Central Australia has been published (Wilson, 1959, *a* and *b*), and that from Western Australia will be sent to press shortly. The tie-line trends (optically derived) of all of these pyroxenes have been determined and are set out in Table 4 as " intersection data ", and Table 5 deals similarly with the chemically derived tie-lines. Where a tie-line intersects the Mg–Ca line at (say) Mg5 Ca95, the symbol XMg5 is used; where one intersects the Fe–Ca line at (say) Fe5 Ca95, the symbol XFe5 is used.

Examples of Co-Existing Pyroxene Pairs

(1) *Central Australia.* (*a*) *Charnockitic granulites.*—Text-fig. 1 and Table 4 show that the optically derived tie-lines of the pyroxenes of the basement charnockitic granulites intersect the Mg–Ca line at about 15½ Mg (i.e. XMg15½)—but note that, due to the use of different optical correlation charts, the intersection may be put at 11 Mg (Wilson, 1959*a*). The sketch map in Text-fig. 1 shows that there is little change of XMg, notwithstanding the scattered locations of the granulites and their varying proximity to major intrusions.

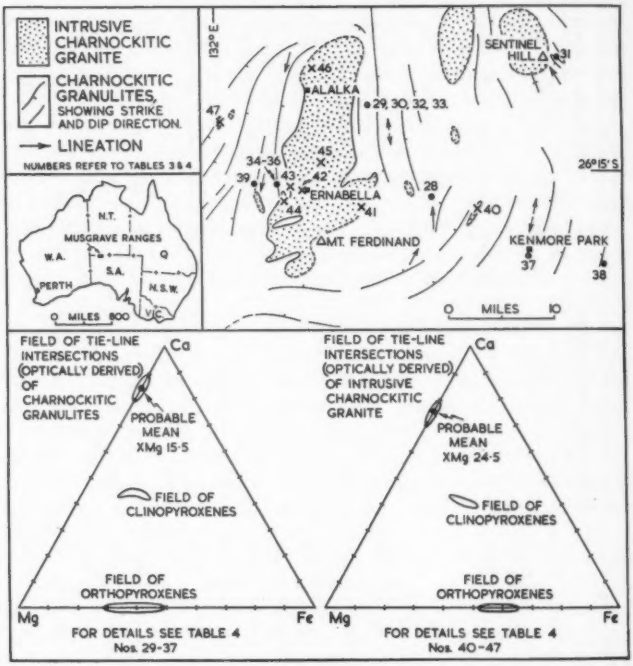
The data from Nos. 29, 30, 32, and 33 (all from one location) indicate that the remarkably consistent XMg value (average 14¼) is not affected by the presence or absence of brown hornblende. However, slightly higher XMg values (about 20) are obtained from the two rocks studied from the eastern Musgrave Ranges (Nos. 31 and 37). It may be significant, from the point of view of geothermometry, that the hornblende in these two rocks has $\gamma = 1{\cdot}674$ and $1{\cdot}675$ (respectively) and is somewhat more green (for γ) than the brownish hornblendes (with slightly higher refractive indices) further west. It should be pointed out, however, that the whole area under discussion in Central Australia is small, for the rocks from which the pyroxene pairs were studied were collected from localities at the most only 30 miles apart. Thus, metamorphic variation is less likely to be important than in India where the charnockitic rocks from which pyroxene pairs were studied were collected from localities up to 350 miles apart.

Only two anomalies have been noticed among the charnockitic granulites. No. 38 is composed of scapolite ($\omega = 1{\cdot}590$), brown

TABLE 1.—OPTICAL DATA AND COMPOSITION (DERIVED OPTICALLY AND CHEMICALLY) OF CO-EXISTING METAMORPHIC PYROXENES LISTED IN TABLE 3

Rock No.	α	β	γ	2V obs.	2V calc.	Composition—Orthopyroxene				Composition—Clinopyroxene								
						from γ	from chem.			from γ, 2V			from β, 2V			from chem.		
						Fe	Ca	Mg	Fe	Ca	Mg	Fe	Ca	Mg	Fe	Ca	Mg	Fe
1	1·707	—	1·722	− 50°	—	46	3	47	50									
1	1·696	1·701	1·724	+ 55°	+ 50½°					44	32	24	44½	32	23½	43	35	22
2	—	1·694	1·699	− 63°	—	27½	?4	73	23									
2	1·690	1·695	1·718	+ 46°	+ 50½°					38	40	22	38	41	21	?38	48	14
3	1·701	—	1·717	− 59°	—	42	1	57	42									
3	1·694	1·700	1·722	+ 59°	+ 55½°					53	27½	19½	53	27½	19½	45	37	18
4	—	1·683	1·688 calc.	− 84°	—	18½	1	83	16									
4	1·686	1·691	1·714	+ 55°	+ 50½°					45½	38	16½	45½	39½	15	47	45	8
5	1·696	1·705	1·708	− 58°	− 59½°	35	1·8	59·7	38·5									
5	1·690	1·695	1·720	+ 50°	+ 49°					40½	37	22½	41	39	20	44·6	39·2	16·2
6	1·705	1·718	1·722	− 54°	− 57½°	46	1·5	53·5	45·0									
6	1·698	1·702	1·722	+ 52°	+ 48½°					41½	35	23½	41	33	26	43·5	35·8	20·7
7	1·720	1·732	1·735	− 56°	− 53°	56½	0·7	42·8	56·5									
7	1·695	1·700	1·720	+ 50°	+ 53½°					40½	37	22½	40	35½	24½	43·0	31·6	25·4
8	1·676	1·684	1·690	− 73°	— 81½°	20½	0·1	75·9	24·0									
8	1·689	1·694	1·711	+ 58°	+ 57½°					52½	35½	12	52	33	15	48·9	41·0	10·1
9	1·725	1·736	1·743	− 58°	− 76½°	63	3·4	34·5	62·1									
9	1·705	1·712	1·732	+ 57°	+ 61½°					46	23	30½	46	22	32	43·7	25·0	31·3

hornblende ($\gamma = 1{\cdot}683$), the two pyroxenes and plagioclase. The scapolite is in granules and appears to be in equilibrium with both hornblende and the pyroxenes. Since this rock is mineralogically unique among the basement granulites, it is not surprising to find that its XMg value (26) is "anomalous". The other "anomaly"



TEXT-FIG. 1.—Tie-line trends (optically derived) of co-existing pyroxenes of intrusive charnockitic granites differ from those of associated charnockitic granulites. The sketch map shows the location of the pyroxene-bearing rocks in the Musgrave Ranges, Central Australia.

(XMg3·5) is shown by a coarse pyroxene amphibolite (No. 39). It is in a roof pendant above a major igneous intrusion and contains an unusual pale green clinoamphibole ($\gamma = 1{\cdot}664$, 2V (+) near 90°) and phlogopite. The clinopyroxene is corroded by orthopyroxene and amphibole. I suspect that the XMg value for both of these rocks is a reflection of their thermal history.

(*b*) *Intrusive charnockitic rocks.*—XMg24½ for the intrusive acid charnockitic rocks contrasts sharply with XMg15½ for the basement

TABLE 2.—OPTICAL DATA AND COMPOSITION (DERIVED OPTICALLY) OF CO-EXISTING METAMORPHIC PYROXENES LISTED IN TABLE 3

Rock No.	γ	2V obs.	Composition O′-pyrox. from γ Fe	Composition C″-pyrox. from γ, 2V Ca	Mg	Fe
10	1·731	− 41° to − 46°	53½			
10	1·719	+ 55°		45	35	20
11[1]	1·737	− 42° to − 52°	58			
11	1·722	+ 56°		46½	31½	22
12	1·737	− 44° to − 51°	58			
12	1·725	+ 57°		48	28½	23½
13[3]	1·734	− 56°	55½			
13[3]	1·726	+ 60°		54½	23	22½
14[4]	1·728	− 43°	51			
14[5]	1·715	+ 50°		41	40½	18½
15	1·730	− 52°	52½			
15	1·719	+ 56½°		48	33	19
16	1·729	− 52°	52			
16	1·719	+ 52°		42½	36½	21
17	1·728		51			
17	1·713	+ 55°		45½	38	15½
18	1·730	− 56°	52½			
18	1·717	+ 50°		41	38½	20½
19	1·762		78			
19[6]	1·717(β)	+ 52°		37½	21½	41
20	1·745	− 64½°	64½			
20	1·720	+ 51°		41½	36½	22
21	1·739	− 63°	60			
21	1·720	+ 50°		40½	37	22½
22	1·734		55½			
22[7]	1·735	+ 49°		36½	26	37½
23	1·683		14½			
23	1·708	+ 58°		52½	38	9½
24	1·685		16			
24	1·709	+ 59°		54½	36	9½
25	1·689		19½			
25	1·709	+ 56°		48	40½	11½
26	1·695		24½			
26	1·699	+ 60°		55	43	2
27	1·711	− 57½°	37			
27	1·728	+ 49[2]		38	31½	30½
28	1·715		40½			
28	1·718	+ 53°		43½	36½	20
29	1·709	− 59°	35½			
29	1·714	+ 55°		45½	38	16½
30	1·709	− 64°	35½			
30	1·715	+ 53°		44	38½	17½

Rock No.	γ	2V obs.	Composition O′-pyrox. from γ Fe	Composition C″-pyrox. from γ, 2V Ca	Mg	Fe
31	1·700	− 68°	28½			
31	1·708	+ 52°		43	44	13
32	1·716	− 54½°	41			
32	1·718	+ 54½°		44½	35½	20
33	1·717	− 55°	42			
33	1·718	+ 55°		45	35½	19½
34	1·720	− 54°	44½			
34	1·721	+ 52°		42	35	23
35	1·708	− 61°	35			
35	1·715	+ 52½°		43½	38½	18
36	1·727	− 53°	50			
36	1·723	+ 52°		42	34	24
37	1·713	− 54½°	38½			
37	1·715	+ 53½°		43½	39	17½
38	1·718	− 59°	43			
38	1·713	+ 57°		50	36	14
39	1·702	− 72°	30			
39	1·712	+ 52½°		43½	40	16½
40	1·735		56½			
40	1·725	+ 51°		40½	33	26½
41	1·727	− 52°	50			
41[8]	1·699(β)	+ 47°		38	37½	24½
42	1·745	− 59°	64½			
42	1·726	+ 51°		40½	32½	27
43	1·733	− 54°	55			
43	1·723	+ 53°		43	33	24
44	1·729		52			
44	1·723	+ 52°		42	33½	24½
45	1·742	— 58°	62			
45	1·729	+ 52°		40½	29½	30
46	1·729	− 51°	52			
46	1·720	+ 53°		43	35½	21½
47	1·746		65½			
47	1·730	+ 51°		39	29	32
48	1·692	− 69°	22			
48	1·712 to 1·716	+ 44°		37 37	45 42	18 to 21
49[9]	1·690	− 76°	20½			
49[10]	1·718	+ 60°		55½	28	16½
51	1·683	+ 87°	14½			
51[11]	1·679(β)	+ 55°		45	48½	6½

[1] 2V (calc.) = − 58°. [2] 2V (calc.) = − 82½°. [3] 2V (calc.) = + 72°; compos. from β, 2V = Ca53½ Mg21 Fe25½. [4] 2V (calc.) = − 59¾°. [5] 2V (calc.) = + 74°; compos. from β, 2V = Ca40½ Mg37½ Fe22. [6] Compos. from β, 2V. [7] 2V (calc.) = + 67; compos. from β, 2V = Ca36 Mg25 Fe39. [8] Compos. from β, 2V. [9] 2V (calc.) = − 76½°. [10] 2V (calc.) = − 80° (badly inconsistent with R.I.'s); compos. from β, 2V = Ca53 Mg21½ Fe25½. [11] Compos. from β, 2V.

Rock No.	*Name, Locality and some Petrologic Notes*
1	Hypersthene diopside plagioclase hornfels, Oslo District, Norway (Muir and Tilley, 1958, p. 404, O–O′). No exsolution lamellae in pyroxenes; no olivine.[3,4]
2	Metamorphosed picrite basalt, ejected block from Kilauea, Hawaii [Muir and Tilley, 1958, p. 404, X–X′ with modification (after personal communication) of the published assumed composition of O′pyroxene from chemical data]. No exsolution lamellae in pyroxene; olivine (Fa 19); minerals possibly in non-equilibrium state; analysed pyroxenes not wholly free from respective associated pyroxene.
3	Hypersthene diopside plagioclase granulite, Scourie, Scotland (Muir and Tilley, 1958, p. 404, S–S′).[1,4]
4	Olivine hypersthene pyroxenite, Scourie, Scotland (Muir and Tilley, 1958, p. 404, T–T′). Olivine (Fa 11); minerals possibly in non-equilibrium state.[1,3,4,5]
5	Intermediate charnockite, Salem District, India (Howie, 1955, pp. 754 and 757, No. 2270).[3,4,7]
6	Basic charnockite, Nagaramalai, Salem District, India (Howie, 1955, pp. 754 and 757, No. 2941).[1,3]
7	Basic charnockite, Pallavaram, Madras State, India (Howie, 1955, pp. 754 and 757, No. 4642A).[1,3]
8	Ultrabasic charnockite, Pammal Hill, Madras State, India (Howie, 1955, pp. 754 and 757, No. 3709).[1,3,5]
9	Intermediate charnockite, Tinnevelly District, Madras State, India (Howie, 1955, pp. 754 and 757, No. 115).[3,4]
10	Basic charnockite, Suomusjärvi, Finland (Parras, 1958, pp. 97 and 104, No. 46).[3,7]
11	Intermediate charnockite, Vihti, Finland (Parras, 1958, pp. 97 and 104, No. 12).[3,4,7]
12	Basic charnockite, Sammatti, Finland (Parras, 1958, pp. 97 and 104, No. 44).[3,4,7]
13	Basic granulite, Koddigvärri, Finnish Lapland (Eskola, 1952, p. 152, Nos. 25 and 27).[3,7]
14	Basic granulite, N. of Merijärvi charnockite pluton, Turku, Finland (Hietanen, 1947, p. 1038, No. 837).[3,4,7]
15	Basic charnockite, C. Naturaliste area, S.W. Australia. 39407.[3,6,7]
16	Basic charnockite, C. Naturaliste area, S.W. Australia. 39409.[3,6,7]
17	Basic charnockite, C. Naturaliste area, S.W. Australia. 39410.[3,6,7]
18	Basic charnockite, C. Naturaliste area, S.W. Australia. 39411.[3,6,7]
19	Intermediate charnockite, C. Naturaliste area, S.W. Australia. 39412.[3,6,7]
20	Intermediate charnockite, C. Naturaliste area, S.W. Australia. 39413.[3,4,6,7]
21	Intermediate charnockite, C. Naturaliste area, S.W. Australia. 39415.[3,4,6,7]
22	Garnetiferous basic granulite, Pallavaram, Madras State, India (Howie and Subramaniam, 1957, p. 569, No. Ch 199).[4,7]
23	Eclogite, Sittampundi Complex, Madras State, India (Subramaniam, 1956, p. 355, No. C 41).[1,3]
24	Bronzitic pyropite, Sittampundi Complex, Madras State, India (Subramaniam, 1956, p. 355, No. C 60).[1,7]
25	Websterite pegmatite, Sittampundi Complex, Madras State, India (Subramaniam, 1956, p. 355, No. C 20).
26	Websterite, Sittampundi Complex, Madras State, India (Subramaniam, 1956, p. 355, No. C 43).
27	Garnetiferous basic granulite, Red Ochre Well, Musgrave Ranges, C. Australia (Wilson, 1959*a*, No. 30543).[1]

Rock No.	*Name, Locality and some Petrologic Notes*
28	Intermediate charnockite, Gilpin's Well, Musgrave Range, C. Australia (Wilson, 1959*a*, No. 30581).[1]
29	Basic charnockite, Taljaritja, Musgrave Ranges, C. Australia (Wilson, 1959*a*, No. 30638).[1]
30	Basic charnockite, Taljaritja, Musgrave Ranges, C. Australia (Wilson, 1959*a*, No. 30639).[1,3]
31	Ultrabasic charnockite, Sentinel Hill, Musgrave Ranges, C. Australia (Wilson, 1959*a*, No. 30230).[1,3]
32	Basic charnockite, Taljaritja, Musgrave Ranges, C. Australia. No. 30637.[1,3,6]
33	Intermediate charnockite, Taljaritja, Musgrave Ranges, C. Australia. 30640.[1,6]
34	Basic charnockite, Mt. Carruthers area, Musgrave Ranges, C. Australia. 30840.[1,3,6]
35	Basic charnockite, Mt. Carruthers area, Musgrave Ranges, C. Australia. 30847.[1,3,6]
36	Enderbite, Mt. Carruthers area, Musgrave Range, C. Australia. 30838.[1,3,4,6]
37	Basic charnockite, Kenmore Park Homestead, Musgrave Ranges, C. Australia. 30553.[1,3,6]
38	Scapolite-bearing basic granulite, 7 miles E. of Kenmore Park Homestead, Musgrave Ranges, C. Australia. 34616.[1,3,6]
39	Augite-bytownite-hornblende-bronzite rock, W. of Mt. Carruthers, Musgrave Ranges, C. Australia. 30855.[1,3,4,6]
40	Ferrohypersthene adamellite, Spinifex Hill, Musgrave Ranges, C. Australia (Wilson, 1959*a*, No. 30558).[2,3]
41	Ferrohypersthene adamellite, Ombagunda, Musgrave Ranges, C. Australia (Wilson, 1959*a*, No. 30385).[2,3]
42	Ferrohypersthene granodiorite, Ernabella, Musgrave Ranges, C. Australia (Wilson, 1959*a*, No. 30789).[2,3]
43	Ferrohypersthene adamellite, Mt. Carruthers, Musgrave Ranges, C. Australia. 30816.[2,6]
44	Ferrohypersthene micro-adamellite, near Mt. Carruthers, Musgrave Ranges, C. Australia. 30871.[1,4,6]
45	Ferrohypersthene adamellite, 4 miles N.N.E. Ernabella, Musgrave Ranges, C. Australia. 30658.[2,3,4,6]
46	Ferrohypersthene granodiorite, near Alalka, Musgrave Ranges, C. Australia. 30476.[2,6]
47	Ferrohypersthene adamellite, Tjakunja, Musgrave Ranges, C. Australia. 30700.[2,3,6]
48	Olivine-bearing pyroxene granulite, Ardnamurchan, Scotland (Muir and Tilley, 1958, p. 404, R–R′). Some exsolution lamellae in clinopyroxene; olivine (Fa 27) is more ferriferous than associated pyroxenes.
49	Ultrabasic charnockite, Pammal, Madras State, India (Howie, 1955, pp. 754 and 757, No. 4645).[2,3,5]
50	Basic granulite, Härkäselkä, Finnish Lapland. (Eskola, 1952, p. 152, Nos. 22 and 26).[4,7]
51	Meta-peridotite, Dawros, Ireland (Rothstein, 1958, No. R).

[1] Pyroxenes contain some fine exsolution lamellae of pyroxene (with or without rods or plates of exsolved iron ores).

[2] Pyroxenes contain coarse exsolution lamellae of pyroxene (with or without rods or plates of exsolved iron ores).

[3] Rock also contains hornblende.

[4] Rock also contains biotite.

[5] Rock also contains spinel.

[6] Number in rock collection of University of W. Australia.

[7] Exsolution phenomena, although not specifically recorded, are likely to be as in other charnockitic granulites.

N.B.—" Charnockite " is used by some authors to name rocks which would be called " pyroxene granulites ", " pyroxene gneisses ", or (possibly) " pyroxene amphibolites " by others.

granulites of the same area (Text-fig. 1). As with the granulites, there is little change of XMg, notwithstanding the fact that samples of the intrusive charnockitic rocks were taken from several distinct and separate igneous bodies (Text-fig. 1).

The origin of the igneous bodies is in doubt. It is likely, however, that they have resulted from deep-seated metasomatic reconstitution of the basement rocks (probably already metamorphosed), and subsequent injection of the rheomorphic crystal mush under physical conditions typical of lower granulite or uppermost amphibolite facies. The presence of sphene and allanite, and absence of orthopyroxene, in the pegmatites from these igneous bodies tend to support this view (Wilson, 1959, *a* and *b*).

(2) *South-western Australia.*—My colleague, Mr. J. G. Kay, has kindly supplied me with the basic data for seven co-existing pyroxene pairs from a very restricted area of granulites at Cape Naturaliste, which is 120 miles S.S.W. of Perth. Much of the area is of high amphibolite facies, but in favourable structural positions rocks of granulite facies occur. A brief note on the regional geology has been published (Wilson, 1958, pp. 64, 79 and map), and Kay expects to publish his detailed study shortly. The area yields tie-line trends averaging XMg 32 (No. 19, determined by β not γ gave XMg21, and is not included in this average).

I would make the tentative suggestion that the high XMg value is a reflection of grade of metamorphism; that is, the charnockitic granulites of the area are low in the granulite facies, whereas those of Central Australia are high.

(3) *Finland.* Four rocks from S.W. Finland (three from within about ten miles of one another near Lohja, Nos. 10, 11, and 12; and one from an area some sixty miles west near Turku, No. 14) give XMg24 $\pm$ 4, and XMg35½, respectively. (Note, however, that the refractive indices of the Turku pyroxene pair are not reasonably self-consistent with observed 2V, hence the estimated composition remains in doubt.) Another pyroxene pair is available from Finnish Lapland, some 600 miles to the north. Although there are large inconsistencies in the optical data, XMg is probably close to 8½.

These data may indicate differences in temperature of crystallization of the granulites between the S.W. and extreme N. of Finland. The impression gained from field and literature studies of the two areas would tend to confirm that the area in S.W. Finland was somewhat below the level of metamorphism of the Lapland area.

(4) *India.* (*a*) *Charnockitic rocks.*—Six chemically analysed pyroxene pairs (Nos. 5–9 and 49) are available from " normal " charnockitic rocks from widely separated areas in Madras State. Critical study of the optical data reveals several inconsistencies which affect chemical

TABLE 4.—INTERSECTION DATA FOR COMPOSITIONAL TIE-LINES (OPTICALLY DERIVED) OF CO-EXISTING METAMORPHIC PYROXENES IN THE SYSTEM "Ca–Mg–Fe"

Rock Group	Rock No.	from γ; XMg	from β; XMg	from γ; XFe	from β; XFe	from γ; probable mean XMg	from γ; probable mean XFe
Norway	1	8	8				
Hornfelses Hawaii	2			16	12·5		
Scotland	48			13 to 20			
Eire	51	19					
Charnockitic granulites (Scotland)	3	1	1				
	4			15	12		
Charnockitic S.W.	10	28					
granulites S.W.	11	25				24	
(Finland) S.W.	12	19·5					
Turku	14	35·5	29				
Lapland	13	8·5	1				
Charnockitic	7	32·5	29				
granulites	8			5·5	11		
(Madras State)	49			14	27½		
	5		5	6			
	6	15	6				
	9	11	6				
Garnet India	22			12·5	18		
granulites Australia	27			24			
Eclogite, etc.	23			5·5			
(India)	24			4·5			4½
	25			3·5			
	26	40					
Charnockitic	15	25					
granulites	16	29					
(S.W. Australia)	17	34·5					
	18	32·5				32	
	19		21				
	20	37					
	21	35					
Charnockitic	29	15					
granulites	30	13					
(C. Australia—	32	13					
see Text-fig. 1)	33	16					
	34	12½				15½	
	35	11					
	36	19					
	28	13·5					
	31	20·5					
	37	20					
	38	26					
	39	3·5					
Intrusive	42	30					
charnockitic	43	23·5					
granites	44	20·5					
(C. Australia)	45	21·5					
	46	27				24½	
	47	23·5					
	41		25·5				
	40	23·5					

composition as deduced from the optical data. Thus, No. 49 has been omitted from Table 1 because, on calculation, the clinopyroxene has 2V (−) about 80° whereas on measurement 2V (+) = 60°. Its chemical analysis, however, is used in Table 5.

The most striking feature of the Indian tie-lines is their erratic distribution when compared with the remarkable consistency of those from both Central and S.W. Australia (see Table 4). It must be remembered, however, that we are comparing metamorphic rocks which occur up to 350 miles apart in India. Thus, if cation replacements

TABLE 5.—INTERSECTION DATA FOR COMPOSITIONAL TIE-LINES (CHEMICALLY DERIVED) OF CO-EXISTING METAMORPHIC PYROXENES IN THE SYSTEM "Ca–Mg–Fe"

Rock group	Rock No.	Mg on Mg–Ca line (i.e. XMg)	Rock group	Rock No.	Mg on Mg–Ca line (i.e. XMg)
Hornfelses	1	25½	Indian charnockitic granulites	5	23
	2	8		6	20½
	48	5		7	22
	51	16		8	14½
				49	21½
				9	15
Scottish charnockitic granulites	3	23			
	4	3			
Finnish charnockitic granulites	13	22½			
	50	27½			

in metamorphic pyroxenes vary in response to temperature change, as I suspect, one would expect this type of wide variation in optical properties and "composition" over such an extensive terrain. Similarly, in Australia, where specimens have been collected from several restricted areas, reasonably self-consistent groups of XMg figures can be expected (as has been found).

(*b*) *Eclogitic and associated rocks.*—Optical data of four pyroxene pairs show that three closely associated rocks of diverse nature (Nos. 23, 24, and 25) have XFe about 4½ (average). Another closely associated rock (websterite, No. 26) has XMg40, which is the most anomalous tie-line of any of the 50 pyroxene pairs studied (Table 4). A factor contributing to this anomaly is that the optical data indicate a clinopyroxene for which the graphs do not cater properly.

It is useful to consider with the eclogitic rocks two other garnet-bearing pyroxene granulites. No. 22 occurs at Pallavaram, near the city of Madras, and No. 27 is from Central Australia. Diopside is

obviously unstable in the presence of garnet in the granulite facies to which both of these rocks belong. It is interesting that of the rocks described in this paper all optically-derived tie-lines of pyroxene pairs in rocks containing garnet (whether of so-called eclogite or granulite facies) cut the Fe–Ca line rather than the Mg–Ca line. Other metamorphic rocks showing evidence of inequilibrium (the olivine-bearing granulites, Nos. 2, 48, and 4) also show intersections on the Fe–Ca line (Table 4).

Possible Causes of "Anomalies" in the Trends of Optically Derived Compositional Tie-lines

Erratic discrepancy between composition determined by optical and by chemical means led Muir and Tilley to make their initial inquiry into the causes of the anomalies. I have listed five possible causes, but in doing so would emphasize that it is most likely that the anomalies shown in this paper are due to a combination of causes.

(1) *Inaccurate graphs to correlate optical data and composition.*— Several workers have pointed out that correlation is good for plutonic igneous orthopyroxene, and the influence of minor element replacement has also been studied (Hori, 1956). Parras has shown, however, that there are significant differences between the optical properties of the orthopyroxenes and clinopyroxenes from metamorphic and from plutonic igneous rocks (Parras, 1958, Figs. 17 and 22). Muir and Tilley (1958, p. 405) note, moreover, that there is a tendency for optical data to indicate an exaggerated iron content for clinopyroxenes of metamorphic rocks. Thus, the over-all effect will be to swing the compositional tie-line clockwise from the "normal" intersection on the Mg–Ca line (about XMg25) toward Ca, and even beyond Ca toward Fe. Again, inaccuracies in the current graphs will be more serious for a Mg-rich pyroxene pair for which the tie-line would normally cut the Mg–Ca line at an acute angle. Indeed, this may be partly the explanation for the fact that the tie-line for the metamorphic ultramafic rocks commonly intersects the Fe–Ca line rather than the Mg–Ca line (e.g. Nos. 8, 49, and 25).

There is an obvious need for graphs to be constructed for the pyroxenes of the granulite facies. Such graphs, however, must be made from a group of rocks from a restricted area, otherwise the effect of varying temperature within the range of the granulite facies is likely to produce erratic results (see (5), below).

An example of the need for reliable graphs to correlate optical data and composition is seen in No. 5 (Table 1). Judging by the consistency of measured and calculated 2V (see also (3) below), the optical data for both pyroxenes appear to be unusually good. However, when

tie-lines are drawn, two diverse trends appear, for there is a significant difference in " composition " of the clinopyroxene as estimated from β and γ, respectively. The tie-line derived from β cuts the Mg–Ca line at XMg5 whereas the tie-line derived from γ cuts the Fe-Ca line at XFe6. This is a cogent reason against trying to compare tie-lines derived from β readings with those derived from γ readings. In this paper, only those from γ are used. Study of data in Table 4 shows the lack of correlation between the β-derived and γ-derived tie-lines of several rocks.

A review of Hori's data concerning the changes in refractive indices and birefringence of pyroxenes with ion substitutions (Hori, 1956, p. 368, Table 7) shows how " anomalies " in the relation between β and γ of a mineral can easily take place. It is clear that the amount of available optical data from actual minerals is as yet insufficient to pursue further the possible use of partial birefringences in metamorphic studies.

(2) *Inaccurate optical measurements.*—It is unlikely that a research man will make gross errors if he observes the normal precautions for optical measurements. However, errors due to faulty refractometer calibration and heating of oils on the microscope stage are easily overlooked. Errors due to heating of the oils are likely to result in consistently high refractive index readings.

(3) *Inconsistent optical measurements.*—Where the three refractive indices and 2V are measured, a calculation of 2V serves as a valuable check on the consistency of the measurements. I suspect that the main cause of inconsistencies is not due to inaccuracy of actual optical measurements, but rather to *variation in composition* of grains in the rock under study. That is, α, β, γ, and 2V all may be accurately measured, but unless single or double variation techniques are used at least four separate grains are likely to be involved in the measurements.

An example of the significance of inconsistent data is in No. 49 (Table 2). The optical data of the orthopyroxene are consistent, but those of the clinopyroxene are very inconsistent. The refractive indices of the clinopyroxene indicate 2V about 80° (negative), whereas the observed 2V is 60° (positive). At least two " compositions " are possible from the optical data, and these give intersections of the two pyroxene tie-lines either at XFe27½ (using β and 2V), or at XFe14 (using γ and 2V). If the data for the co-existing orthopyroxene were similarly inconsistent, it would be more futile to draw the several possible tie-lines for the rock.

(4) *Exsolution phenomena and other inclusions.*—Notwithstanding normal careful hand-picking for chemical analysis of pyroxenes which

contain exsolution lamellae or other small inclusions, the sample is likely to contain many composite or impure grains. Under such conditions the composition from the chemical analysis must differ from that deduced from the optical properties of clean grains in the sample. In rocks of the granulite facies, for instance, exsolution lamellae and other tiny inclusions are very common and exceedingly difficult to remove. This is not a bad thing, of course—a chemical analysis of host + exsolution lamellae is very desirable for some petrologic purposes.

An interesting example of this comes from Madras. Nos. 8 and 49 are pyroxene pairs from closely associated and very similar ultrabasic charnockites. The orthopyroxenes on chemical analysis show 24 and 27 per cent of the Fe component, respectively, yet $\gamma = 1{\cdot}690$ for both. No. 49, however, has plentiful lustrous plate and rod inclusions which are concentrated in the cores of the grains. In No. 8 the plate and rod inclusions are less frequently developed (Howie, 1955, p. 731). It seems to me that the actual orthopyroxene of both rocks is now virtually of the same composition, but that originally No. 49 was richer in Fe ++ and Ti, both of which have subsequently exsolved to form the inclusions.

Exsolution lamellae of pyroxene and needle and plate inclusions are commonly absent from pyroxenes of the hornfelses. In this they differ from most of the charnockitic rocks, for these have a longer period of cooling. I believe that there is insufficient evidence available to say that there is no significant chemical difference between co-existing pyroxenes from the hornfelses and those from the metamorphic rocks of granulite facies. Muir and Tilley (1958) publish chemical data of three hornfelses (Nos. 1, 2, and 48), and another has recently come to hand (No. 51). The tie-line (chemical) for No. 1, the hornfels from Oslo, cuts the Mg–Ca line at XMg25½, which is ideal for an igneous assemblage. However, the other three cut the Mg–Ca line at XMg8, XMg5, and XMg16, respectively (Table 5).

At first sight one may be tempted to assume that these " anomalies " may be dismissed as being due to impurities in the analysed samples. It is true that the analysed samples of both pyroxenes of No. 2 contain impurities of each other, but the " compositions " used to derive the data for Table 5 are those suggested by Muir and Tilley after allowance has been made for this. Again, the range of refractive indices recorded in No. 48 for the clinopyroxene suggests a variable composition for clinopyroxenes at least. However, the possible impurities in the chemically analysed sample are little worse than those of many analysed samples which are accepted for petrologic comparison. Eskola, for instance, suspects admixture of both pyroxenes in the analysed samples of both of the clinopyroxenes from the Finnish charnockitic granulites

(Eskola, 1952, p. 155). Yet these analyses are used by Muir and Tilley (1958, Text-fig. 1) to support their thesis that in metamorphic assemblages the tie-line trends show no significant departure from those recorded for plutonic igneous assemblages.

Many of these comments on the chemical features of the hornfelses apply equally well to the tie-lines derived from optical data.

(5) *Metamorphic history and status.*—As pointed out in several places above, I suspect that a major cause of " anomalous " departure of tie-line trends (derived either chemically or optically) is to be sought in the metamorphic history or status of the rocks concerned. Thus, for the Central Australian rocks the two main tie-line trends, both of which differ from those of metamorphically restricted areas elsewhere in the world, are thought to have geothermal significance.

Moreover, tie-line trends appear to indicate instability relationships in many rocks. Thus, we note that all three olivine-bearing granulites shown in Table 2 of Muir and Tilley (1958) show consistent " anomalies " [tie-line intersections (chemical) on Mg–Ca line for Nos. 2, 48, and 4 (my numbers) are XMg8, XMg5, and XMg3, respectively—see Table 5, this paper]. Another granulite where olivine is probably present (but published data are inadequate to determine its relationship to co-existing minerals) is No. 51; its tie-line (chemical) cuts the Mg–Ca line at XMg16. Muir and Tilley (1958, p. 407) point out that two of the three olivine-bearing rocks (my numbers, 2 and 4) are probably in a non-equilibrium state. Moreover, although for good reasons they consider that the third rock (No. 48) was completely recrystallized, I suspect that the position of the tie-line and the variation recorded in refractive indices of the clinopyroxene both indicate that the three mafic minerals (olivine, orthopyroxene, and clinopyroxene) are not yet in equilibrium. Empirically, it would appear likely that olivine-bearing two-pyroxene metamorphic assemblages are commonly in a non-equilibrium state, and may be expected to contain pyroxene pairs differing significantly from those of plutonic rocks.

Not only is this apparent from a study of tie-lines derived chemically, for the tie-lines trends derived optically show even larger deviation from those of the plutonic igneous types. The tie-lines (optical) of Nos. 2, 48, and 4 cut the Fe–Ca line at XFe16, XFe13 to XFe20, and XFe15, respectively. (Table 4.) This is reminiscent of the intersections of tie-lines of garnet-rich granulites and two-pyroxene eclogites (Nos. 22, 23, 24, and 27) which I believe show some petrographic evidence of a non-equilibrium state (Table 4).

In this connection we can recall that the only two Central Australian granulites which show an " anomalous " tie-line for the pyroxenes area scapolite-bearing granulite (No. 38), and an ultramafic metamorphic rock containing an unusual amphibole (No. 39).

Conclusions

I believe that evidence is insufficient to state that " in metamorphic assemblages carrying two pyroxenes the [chemically derived] tie-line trends show no significant departure from that recorded from plutonic igneous assemblages " (Muir and Tilley, 1958, p. 406). I agree, of course, that optical data are insufficient at present to estimate, either consistently or accurately, the true chemical composition of metamorphic pyroxenes [that is, so far as the three major components (Ca, Mg, and Fe) are concerned]. However, departures from the so-called typical trend of tie-lines for plutonic igneous rocks deserve special attention, whether the tie-lines be derived from chemical or optical data. In this paper I have pointed out some of the possible causes of the departures, but it is obvious that the dearth of reliable chemical and optical data on the co-existing minerals of individual rocks is our greatest hindrance in the study of the problems here discussed. The valuable contribution by Muir and Tilley shows, of course, that they are fully aware of this.

De Vore (1957, p. 183) contends that the Mg–Fe^{2+} distribution between co-existing orthopyroxene and clinopyroxene is strongly controlled by the relative distribution of Al (in six-fold co-ordination), Fe^{3+} and Ti between the two minerals, and by the level of influence that these cations have on the six-fold co-ordinated positions. It is, therefore, of some interest to note that a large proportion of the $Mg/Mg + Fe^{2+}$ " anomalies " in the pyroxenes studied by De Vore were found in metamorphic pyroxenes. Assuming De Vore's conclusions are valid, the fundamental question arises as to the cause of the variable content of Al-6, Fe^{3+}, Ti and other ions in the pyroxenes—that is, is variation due to the metamorphic history of the pyroxene, or is it simply an indication of, say, the total Ti in a rock ?

In this connection it is important to note Hori's conclusions on the effect of these ions on the refractive indices of the pyroxenes (Hori, 1956, p. 369). In both orthopyroxene and clinopyroxene α is decreased by substitution of Si by Al, whereas both β and γ are increased in orthopyroxene and decreased in clinopyroxene. The variations are such that birefringence is increased in both pyroxenes. However, the resultant effect of Al-6 on the birefringence is to raise it for orthopyroxene and to depress it slightly for clinopyroxene. Although α, β, and γ are all raised by substitution of Mg by either Ti or Fe^{3+} in both pyroxenes, the increases are such that birefringence is decreased in both pyroxenes. Variations of these types obviously will affect compositional tie-lines deduced from optical data. Such variations may yet prove so serious and unpredictable that the significance of such tie-line trends is nullified. Before this conclusion is reached,

however, it is desirable to carry out research to discover the effect which temperature (and pressure) conditions have on the relation between the total amount of these components in a rock, and their distribution among the pyroxenes of the same rock.

Our prime need in this connection is to have several pyroxene-bearing metamorphic rocks studied thoroughly. Howie has made an admirable advance towards this goal, for his chemical " dissections " and optical studies (for instance, of No. 2270 (my No. 5)) have filled a long-felt need (Howie, 1955, p. 742, Table 3). However, I am convinced that chemical data are not enough. In studying charnockitic rocks, for instance, the full tectonic history is equally as important as their chemistry—indeed, here as elsewhere, tectonism and chemistry are very closely linked. The over-all dark greasy appearance of charnockitic rocks makes them look so homogeneous in hand-specimen that their normal banded or streaky character is commonly overlooked (Wilson, 1957).

In banded metamorphic rocks of this type, it is not enough to analyse the pyroxenes of the whole rock. There may well be significant differences in the composition of the pyroxenes from layer to layer in the rock. This is not uncommonly found in plagioclase, in which small variations of composition are readily detected by variation in extinction angles. I believe that, although many of us have noticed small but significant variations in the plagioclase, we have failed to appreciate that comparable (but less obvious) variations in composition probably occur in the associated mafic minerals.

I would conclude that, notwithstanding the likelihood of inaccuracies in some of the available data, there is sufficient evidence to indicate the probability that a study of accurate optically derived tie-lines of the pyroxenes of metamorphic rocks will yield valuable relative temperature data.

As a first approximation from available optical data (for rocks with the minerals in equilibrium) it would appear that where high XMg values obtain, the rocks are either mobilized granulites or pyroxene granulites close to the " boundary " of the granulite and amphibolite facies. For pyroxene granulites progressively further removed from the amphibolite facies boundary, XMg may be expected to diminish.

Acknowledgments

I wish to thank Professor R. T. Prider for critically reading the manuscript, and Mr. J. G. Kay for kindly supplying the optical data pertaining to the Western Australian granulites. I must accept full responsibility, however, for the accuracy of the calculations.

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The Old Red Sandstone of Eastern Ekmanfjorden, Vestspitsbergen

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Abstract

The Lower Old Red Sandstone deposits of Ekmanfjorden belong mainly to the Lykta Division, but pale sandstones with *Giganthaspis* representing the Kapp Kjeldsen Division also occur in the core of what is probably a N.-S. striking anticline. The lower beds of the Lykta Division are soft, dull and silty and not very fossiliferous; above them are some 1,100 feet of sandstones and conglomeratic beds alternating with shales or marls. Locally abundant vertebrate remains include the guide fossil *Doryaspis*. Various sedimentary structures suggest rapid deposition on a broad shallow delta front.

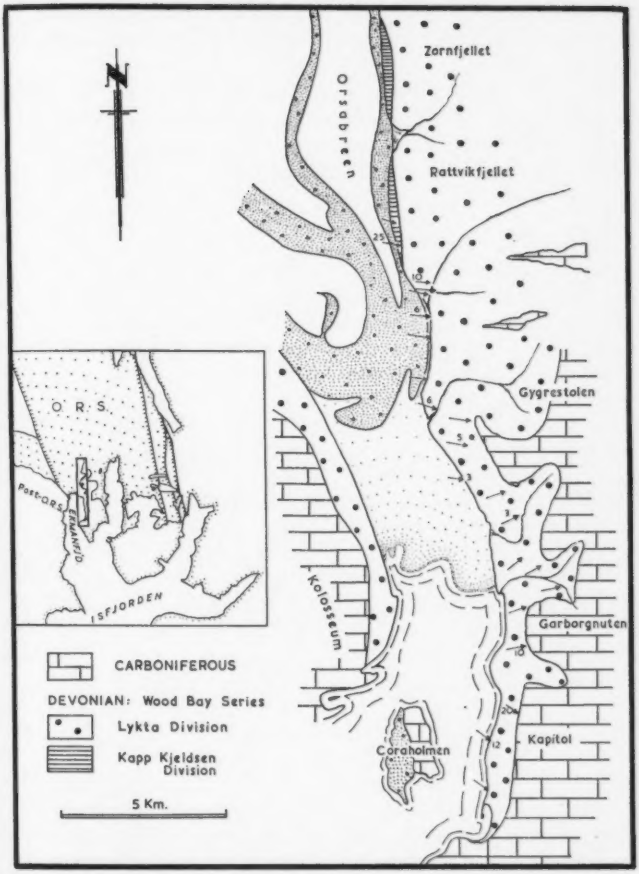
Introduction

In their account of the Old Red Sandstone deposits of Spitsbergen Føyn and Heintz (1943) briefly described the Wood Bay Series in Ekmanfjorden in central Vestspitsbergen and passing reference has also been made to these rocks by Bates and Schwarzacher (1958). The present paper records further details of the lithologies, rock sequence and other aspects of these strata examined during the course of the Birmingham and Exeter Universities joint Spitsbergen Expedition of 1958.

Between Ekmanfjorden and Billefjorden the Old Red Sandstone disappears below an unconformable and southwards-tilted sheet of Upper Carboniferous limestones and is bounded to east and west by strong faults. Ekmanfjorden itself is only a few miles east of the western boundary fault of the great central Spitsbergen graben, a depression which may have been subsiding during Devonian times prior to the major fault-movements of the Upper Devonian. The Wood Bay Series, which is reputed to be up to 6,550 ft. thick, occupies the greater part of this graben and is divisible thus:

3. Stjørdalen Division: guide fossils—*Monaspis*, *Nectaspis*, phlyctaenaspids.
2. Lykta Division: guide fossils—*Actinolepis*, *Doryaspis*.
1. Kapp Kjeldsen Division: guide fossils—*Arctaspis*, *Arctolepis*, *Giganthaspis*.

Of these, only the lower two are present at Ekmanfjorden, the Kapp Kjeldsen Division being confined to a narrow belt on the eastern side of Orsabreen while the Lykta Division appears to outcrop over the remaining ground. More or less continuous exposures are found along the coastline and on the crests of the mountain spurs but the rough,



TEXT-FIG. 1.—Geological sketch map of the outcrops of the Wood Bay Series at the head of Ekmanfjorden, Vestspitsbergen, based on an unpublished topographical map from Norsk Polarinstitutt (by courtesy of Dr. A. K. Orvin).

scree-clad slopes have comparatively few large continuous exposures. Nevertheless it is possible from some distance to interpret the geological structure from the lines of scattered crags.

The English—Norwegian—Swedish Expedition of 1939 visited Ekmanfjorden to gather palaeontological material from the coastal slopes south-west of Gygrestolen. Few of the fossils collected there

have been described, though the work is now in progress, and a complete detailed investigation of the Wood Bay Series vertebrate faunas is still awaited.

The writer is pleased to acknowledge the financial assistance given to the 1958 expedition by the Universities of Birmingham and Exeter and by the Royal Geographical Society. He is also grateful to the friends, especially Mr. P. A. Garrett, who gave such ready assistance in the field and laboratory.

Kapp Kjeldsen Division

Pale green sandstones with *Giganthaspis* appear to characterise the uppermost part of the Kapp Kjeldsen Division over a wide area in north Vestspitsbergen. Føyn and Heintz (1943) record them in Bockfjorden, south and east Woodfjorden and Vestfjorden. Reference to their account leaves no doubt that the beds described below lie within the same part of the Wood Bay Series. The topmost 130 ft. or so of the Kapp Kjeldsen Division crops out on the lower western slopes of Zornfjellet and Rattvikfjellet. Though not extensive, the exposures show that distinctive light green, tough quartzitic or very calcareous sandstones and conglomerates form the bulk of the strata. These beds lie between even and parallel bedding-planes and are up to about 18 ft. thick, being separated by thinner seams of green, sometimes red-blotched clay-shale. (Colour ranges are 10G 6/2 to 5BG 5/2 for greens; 5R 3/4 to 10 5/4 for reds on the N.R.C. Rock-colour Chart.) Less than ten per cent of the examined section of Kapp Kjeldsen beds was red and the reason for the extensive green coloration is not obvious from preliminary examination. Locally in the fossiliferous beds it may be associated with finely divided organic matter which apparently fixes the iron present in the ferrous state. Although some chloritic detritus is present it does not contribute appreciably to the colour.

The detrital mineral assemblages differ little amongst sandstone samples collected at random. Quartz dominates the assemblages and forms about ninety per cent of the clastic fraction. Feldspars are limited to a few laths of albitic plagioclase but there is frequent white mica and a little brown mica and clear green chlorite. Opaque minerals are rare. The grains are all angular or sub-angular, showing little sign of abrasion during transport. On some bedding-planes flakes of white mica are most conspicuous, reaching up to 2·5 mm. in diameter. The flakes are often locally gathered into small patches between which there are only a few very small mica fragments scattered over the rock surface. Presumably this is due to a form of sorting action by the current carrying the sediment. Some of the sandstones have scattered pea-sized ironstone nodules which give rise to local yellow-weathering areas. Calcite cementation is usual (lustre-mottling by calcite being found in

a number of beds); less frequently the cement is quartz and, rarely, it is ferruginous material. The clastic assemblage indicates derivation from metamorphic rocks, probably of a fairly low grade, and transport of the sediment would appear to have been rapid with little time spent in movement over the interface.

Cross-bedding on a small scale (with foreset beds 1 to 3 in. in thickness) was found in some of the strata, being most conspicuous in the lower 10 or 12 in. of the sandstones. The sandstone under-surfaces which were examined revealed no sole structures of interest beyond one or two very minor irregularities which may reflect erosional features cut in the marl or shale beneath. Basal conglomeratic layers, all remarkably uniform in general aspect, occur in some of the sandstones. They are somewhat lenticular, ranging up to 2 ft. in thickness along the strike, and contain well-rounded and fairly well-sorted vein quartz pebbles up to 2 in. in diameter. In another distinct, but less common, variety of conglomeratic bed, the " pellet cornstone " of other Old Red Sandstone regions, large flakes of marl or shale, siltstone and nodules of concretionary limestone or " race " are set in a calcareous sandy or silty matrix. These flakes or pellets should perhaps be termed " intraclasts " since they appear to be derived from the erosion of local Devonian sediments. Folk (1959) says " Intraclasts may be torn up from sedimentary layers almost immediately after they have been laid down, or under more severe conditions may be produced by erosion of layers that have become buried some feet below the sea floor." (The term " pellet " is now commonly used for small bodies of microcrystalline calcite ooze which may be of an invertebrate faecal origin).

Ripple-marks and desiccation-cracks were not found. Circular structures up to 4 in. in diameter and similar to the " spring-pits " described by Quirke (1930) were encountered on the upper surface of one thick sandstone. In relief they reach an inch or a little less and eight of them were found scattered over about an area of about three feet square. Quirke's specimens were up to 2 ft. across and were said to be formed in littoral sands by ascending waters. There is no direct evidence of a littoral character in the present sandstones but as part of a deltaic formation it is possible that upwelling waters disturbed them, though whether they mark springs above or below low water mark is not known. Identical, and also larger, structures have been found in the Old Red Sandstone of Britain (Thomas, 1957). That they mark springs produced as a result of compaction squeezing out pore-fluids from the sand body seems possible.

Jointing in the sandstones on Rattvikfjellet is remarkably regular, breaking up the beds into blocks about three feet cube. In the higher beds of the Wood Bay Series to the south it is much less regular, but

throughout the Series a strong joint system persists, aligned approximately north-south, i.e. parallel to the fold axes. Jointing on E.-W. lines is not so conspicuous.

Fossils collected *in situ* on scree or moraine on Orsabreen include large pteraspid plates and other fragmentary vertebrate remains. The best preserved specimens, among which are almost complete large pteraspid dorsal and ventral discs, probably of *Giganthaspis*, are found in the finer intraclast conglomerates and sandstones. The largest disc is 12 cms. long and 8 cms. broad while the smallest fragment is more than half this size. These fossils are gathered together in loose clusters or groups, usually with concave sides uppermost, in sandstones which are otherwise quite barren. Arctolepid and ther very stout arthrodire fragments occur but are generally too broken to be named: ?*Arctolepis sp.* is tentatively identified. From a loose block containing ?*Giganthaspis* a single very large porolepid plate was also recovered. In the conglomeratic rocks the fossils are less well preserved, having suffered during vigorous transportation. As a rule they are also fairly well sorted in size or weight, with broken portions of large plates predominating; small complete plates or scales were not found.

Fragmentary plant remains occur as rare thin carbonized films in the conglomeratic bands. None was collected.

On several bedding planes a few smooth " trails " an inch or a little less in width, half an inch deep, and up to nine inches long occur. They are usually curved or horseshoe-shaped and are most probably of organic origin. (See text-fig. 5.) Trails of the type figured by Holtedahl (1926) were not seen.

Lykta Division

South and east of Rattvikfjellet, and probably north of Zornfjellet too, the outcropping rocks belong to the Lykta Division, giving rise to red-brown hillsides with occasional crags or bands of green sandstone or conglomerate. The junction of the soft red sandstone of the Lykta Division with the underlying harder green more massive sandstones of the Kapp Kjeldsen Division here is sharp and apparently conformable. Two broad subdivisions may be distinguished in the Lykta beds on the slopes above Ekmanfjorden. The lower is typified by dark and dull fine-grained silty sandstones and shales, about 250 ft. thick; the upper shows a sharper division between marls or shales and sandstones and locally reaches about 1,100 ft. in thickness. The same fossils were found in each subdivision but the upper appears to be the more fossiliferous.

The lower subdivision

This part of the Lykta Division was briefly examined in the low

ground between Leksandfjellet and Rattvikfjellet. It presents a sequence of silty, muddy sandstones of a dull red colour (about 10R 3/4,), occasionally cross-bedded and up to 20 ft. in thickness. A ratio of finer-grained beds to sandstones of about 3:2 was estimated in the field, and the mountainsides here do not have the conspicuously stepped and rough appearance of those where the higher beds outcrop. Green spots and locally green bedding planes and joint planes are common. White mica is abundant in nearly all grades of sediment but is often so finely divided that it is not as immediately conspicuous as in higher beds. The shales and clay beds between the sandstones are often finely laminated and somewhat silty and may alternate rapidly with thin courses of silty sandstones. Unlike the marls of the upper part of the Lykta Division, they contain only rare thin bands of small calcareous nodules.

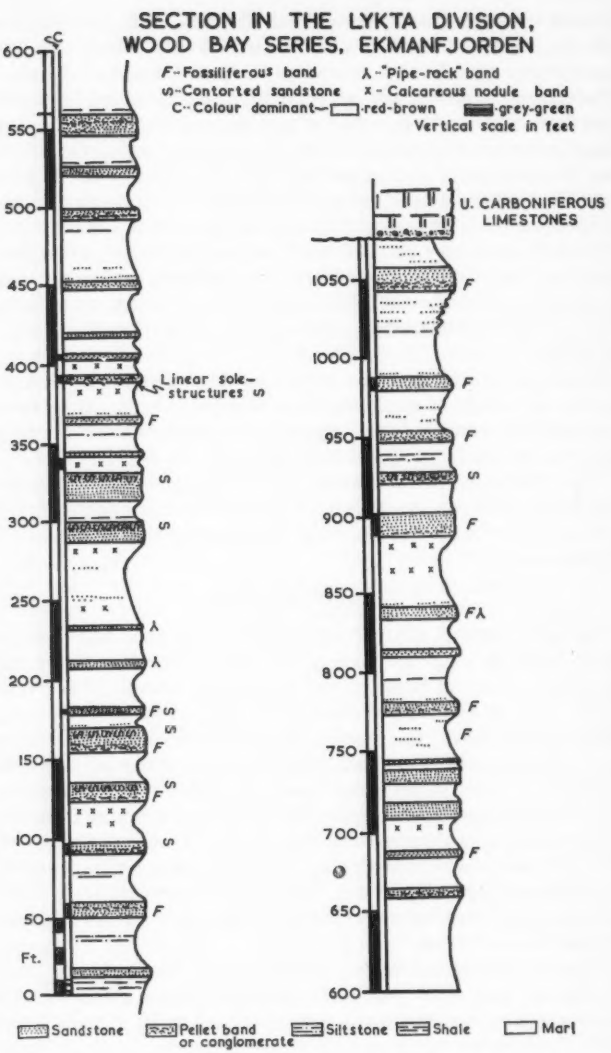
Intraclast beds are common and include lenticular bands consisting almost exclusively of marl flakes several inches across. These bands are rarely calcareous and fossils are not common. Pebble conglomerates like those of the Kapp Kjeldsen Division were not found.

The collection of fossils from this group of strata includes *Doryaspis*, porolepid scales and a single very large cephalaspid cornual piece. Neither *Giganthaspis* nor any other large pteraspidomorph seems to be present.

The upper subdivision

The sandstones in this part of the Series vary from fine to coarse grade and from tough and indurated to soft and friable. Calcite and ferruginous material make up the cement and lustre-mottling often results. Almost all the sandstones are micaceous, often with very large flakes of white mica. They contain also much strained quartz and fine green chlorite, and some beds have large quantities of interstitial argillaceous paste. A few green bands are remarkably persistent and can be traced from one mountain side to another, but more frequently the green coloration is very local. The sandstones are usually sharply defined by their basal bedding planes but grade imperceptibly upwards into siltstones and silty shales. Thicknesses approach 18 ft. Thin courses of silty sandstones are common in the major marl or clay units between the sandstones.

Intraclast conglomerates are numerous and are by far the most fossiliferous beds, occurring at the base of the thicker sandstones and also as irregular lenticles in both sandstones and marls. Prominent and richly fossiliferous examples were found at the mouth of the valley draining the south-western side of Gygrestolen. Here they are up to 5 ft. thick and are fairly uniform in composition and grade throughout. Field observation suggests that the grading or sorting coefficient



TEXT-FIG. 2.—Diagrammatic section of the upper part of the Lykta Division outcropping on the ridge running south-south-west from Gygrestolen.

is high, with usually rounded flat pieces of marl, shale or even siltstones up to six inches in diameter. Calcareous nodules may form an appreciable part of the assemblage in some beds. The matrix is a more or less ferruginous sand with a calcite cement and the intraclasts are usually separated from one another by varying amounts of matrix. At some levels the intraclast material within the marls is gathered into a sheet of thin more or less separate lenticles resembling gravel-filled channels between 2 and 4 ft. across and up to 1 ft. thick.

Cross-bedding is common in sandstones and conglomerates at all levels. The sandstones may also show both current and oscillation types of ripple-marks and some of the finer sandstones are extensively ripple-bedded. Perfect small desiccation-crack polygons, spread over large areas at many levels, indicate frequent sub-aerial exposure and emphasize the prevailing extreme shallowness of deposition. Two features of special interest in the sandstones are the extensive development of contorted bedding and an occurence of strong linear sole-markings.

Contorted bedding

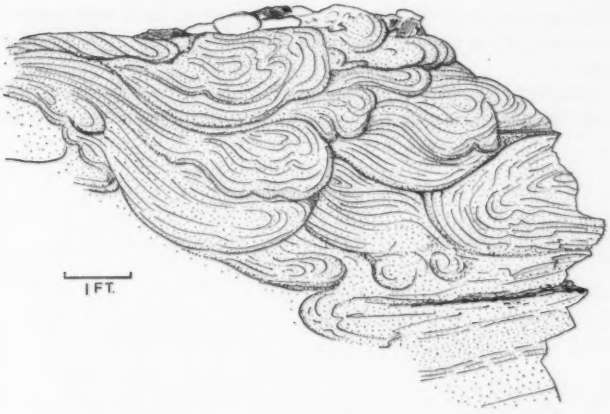
Sandstones showing contortions of their internal laminae were recorded at several points in the section south-west of Gygrestolen and also occur elsewhere. A typical example shows about 6 ft. of fine- to medium-grained sandstone with normal cross-bedding passing up into an intensely contorted mass of about the same or slightly greater thickness. The upper part of the stratum is built up of tightly packed generally pillow-shaped convolutions or " rolls " up to 2 ft. thick and 3 ft. or more along their long axes. In cross section they are usually flattened rather than round, so that " balls " similar to those described in strata elsewhere were not found. The sand laminae within each roll appear to be separated by paper-thin muddy films, not easily discerned in the fresh rock. Each roll or unit is in appearance a distinct and discrete entity with the laminae sharply recurved or " wrapped " around the centre of the roll and sharply truncated over part of its surface. The higher units are pressed and moulded about the rounded upper surfaces of the lower but there is no slickensiding, polishing or other sign of tectonic disturbance between the units, and a sedimentary origin is clearly indicated. While the lowest often incompletely convoluted units disturb the laminae beneath them to a depth of several inches, the upper surface of the contorted mass is, locally at least, relatively flat. In places there are signs of erosion prior to the deposition of overlying materials. On the thickly scree-covered mountain sides it was not possible to trace the contorted bands far from the crest of the ridge, but over a distance of 25–30 yards they showed no signs of diminishing. Brief examination of these exposures revealed no

apparent linear arrangement to the axes of these convolutions. Similar seemingly chaotic structures are recorded in Pembrokeshire and are regarded as produced by turbulent flow of sediment (Keunen, 1948). This must have taken place very shortly after deposition, the sand being slightly muddy and in a hydroplastic condition. Movement of material of this kind may take place down gradients of a degree or two and may be triggered off when a critical gradient or local thickness is reached. This may have been produced by differential compaction of underlying sediments or by an earth tremor. Denudation of the Hecla Hoek highlands to the west is generally held to have produced a series of broad and rapidly developing delta fans in central Spitsbergen. The outer slopes of these fans would be likely places in which to find subaqueous slumping and slipping of water-filled and rapidly laid masses of sand. From time to time, no doubt, movement occurred along the fault lines bounding the central Spitsbergen graben, perhaps assisting to dislodge the newly accumulated deltaic fans or sheets. The lowest convolutions in such a sheet seem to be less tightly rolled and crumpled than the higher and hence seem to have travelled only a little way: the higher units appear to have moved further and become completely separated from their source. The size of the area over which such movement might have occurred remains undetermined but it seems likely that it may have been extensive. The alternation of the major fine and coarse grade units in the Red Bay and Wood Bay Series is typical also of other Old Red Sandstone formations and may reflect a depositional response to continuing mild tectonic activity. Fault tremors may or may not have been important or frequent, but continued intermittent uplift of the source area of the Old Red sediments is often indicated.

Linear structures

Small sole-markings are present on many of the sandstones and pellet beds. They almost invariably seem to represent infillings of erosion surfaces, though a post-depositional tectonic origin may in some instances be possible. At only one horizon on the undersurface of a medium-grained ripple-marked and current-bedded sandstone were continuous linear structures observed. They occur as two sets of distinct ridges, the stronger set consisting of more or less uniform, parallel-sided ridges of almost semi-circular outline and 1 to 2 in. across. Each ridge is continuous over the entire exposed surface of the rock, i.e. 8 ft. square or more, and is separated from its fellows by about 1 ft. to 15 in. This set of markings seems to be superimposed upon a fainter series of somewhat discontinuous ridges or "crinkles," having the appearance of a strong primary lineation. There is, however, an angle of about 15 degrees between the two sets

of markings, and their genetic relationships to one another remain obscure. The fainter set may indeed be a primary lineation, though it is stronger than is usual in Old Red Sandstone deposits. The heavier markings seem to be post-depositional, imposed upon the fainter and resembling the groove casts recorded by other writers (see, for example, Shrock, 1948). No satisfactory explanation for them has yet been found.[1] Primary lineation occurs in many other sandstones.



TEXT-FIG. 3.—Diagram to show contorted bedding in a thick sandstone in the upper part of the Lykta Division, ridge south-south-west of Gygrestolen. (From a photograph.)

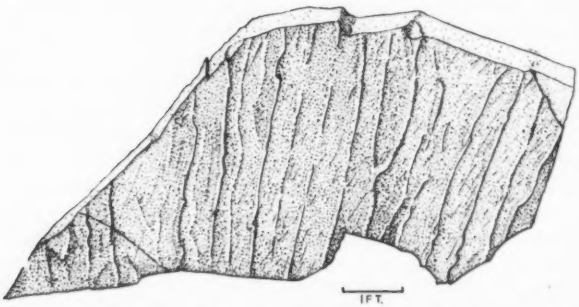
Markings possibly of organic origin

No undisputed trace-fossils of the *Chondrites* type have been discovered, though they are recorded in similar rocks elsewhere in Spitsbergen. Two types of marking, however, were found at many levels in the Lykta Division. These are pipe or burrow-like vertical structures resembling a letter Y in shape and the horizontal horseshoe-like features found also in the Kapp Kjeldsen Division. The vertical

[1] In a paper published since the above was written K. Jingwha Hsu (Flute- and Groove-Casts in the Pre-Alpine Flysch of Switzerland, *Amer. Journ. Sci.*, cclvii 529–36) suggests that drag marks—which are essentially similar to the features described above—are produced by a sediment-water mass moving at a rate at which the coarsest load is being deposited. The basal stratum of coarse sediment is forced to continue its forward motion by the moving mass above and to assume laminar flow. Excavation by this current is thus confined to linear channels, and successive currents moving in different directions may superimpose differently aligned sets of grooves. Though this hypothesis concerns greywackes, a similar explanation may hold in the present case.

markings are pipe-like bodies in the sandstone, bifurcating in their upper half. They may be as much as 15 in. long and an inch in diameter and terminate as a faint round impression on the upper surface of the sandstone. In side view they are marked by a slight purplish or greenish colour change and there is some vague and almost imperceptible disturbance of the material within the pipe. Frequently the pipes contain a little clayey, red material almost as a skin around the sandy filling. It is the bifurcation of the burrow or pipe that is puzzling. The animals producing the structures appear to have been numerous for many of the beds invested with them have six or more per square foot.

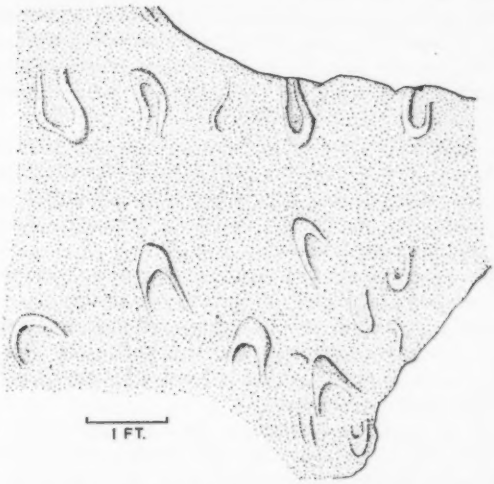
Reference has been made above to the lines of crags which project from the mountain sides. They have been scrutinised along the coastal



TEXT-FIG. 4.—Diagram to show linear markings on the under surface of a sandstone in the upper part of the Lykta Division, ridge south-south-west of Gygrestolen. (From a photograph.)

(strike) sections for two reasons: (i) to determine the geological structure and (ii) to estimate if possible the lateral extent of the individual sandstone and conglomerate units which produce them. Distinctively coloured or otherwise conspicuous beds lend themselves best to this study, and the impression gained from the bright green beds occurring at several levels is that they and the other crag-forming beds are continuous, and over a distance of several miles retain a remarkably uniform thickness. Some of them, especially those with slumped bedding, might be used as marker horizons for further studies. The individual intraclast conglomerate layers and lenses within these sandstone units are, however, very probably highly discontinuous. In their continuity and unchanging thicknesses these sandstone units are more like those of the Lower Old Red Sandstone of South Wales (Pembrokeshire) than those of Shropshire where the units vary greatly

in thickness over comparatively small areas. This change in habit between coeval rocks of the two areas must reflect differing conditions of accumulation, a topic to be discussed elsewhere by the writer. The general aspect of the succession is deltaic with deposition over a shallow, periodically exposed and inundated delta front of fairly broad extent. The sandstone and conglomerate units rarely exceed 25 feet and never exceed 30 feet in thickness but the marl or shale units may be 50 feet or



TEXT-FIG. 5.—Diagram to show horseshoe-shaped markings, possibly of organic origin, in a sandstone in the upper part of the Lykta Division, valley south of Gygrestolen. (From a photograph.)

more, giving a total ratio of sandstone to marl of about 1:3 in the measured section. Deposition of the sandstone conglomerate units probably took place rapidly in response to periodic tectonic or climatic events or both. Fairly thorough chemical weathering of the source area appears to be indicated by the red marls which were perhaps transported comparatively slowly into the depositional basin.

Palaeontology

Dominating the fauna of this part of the Wood Bay Series are *Doryaspis*, porolepids, and phlyctaenaspids and they are most commonly encountered in the conglomerates. They are less numerous but often are better preserved in the sandstones. In many horizons they have become impregnated and coated with a thin film of fine red haematite.

Doryaspis, named as the guide fossil for the Lykta Division by Føyn

and Heintz (1943, p. 15), is abundant at all levels, occurring alone or with other vertebrates. Slabs of fine sandstone with twenty or more weathered and closely gathered dorsal and ventral discs (and other fragments) were found in the screes. In the conglomerate the cross-bedding usually inhibits the parting of the rock into slabs with a large surface area but similar numbers of fossils may be present on bedding surfaces of comparable size. Current-sorting has evidently taken place, producing local concentrations of fossils and the faunules bear every sign of being thanatocoenose accumulations produced after the break up of the animals' carapaces. Though no *Doryaspis* was found intact during the present work in Ekmanfjorden, remarkably complete specimens have been obtained previously and are at present being studied by Miss N. Heintz in Oslo.

Isolated porolepid scales are very abundant here and there throughout the Lykta Division and other parts of the skeleton are known in a fragmentary condition. Again, the concentration of separate scales and bone fragments into local patches suggests a vigorous transport of the hard parts of the animals following death.

The arctolepids are also represented by fragments noticeably more common and larger in the pellet beds than in the sandstones. This may be because their generally heavier construction does not render them as prone to *post mortem* damage as are the more fragile *Doryaspis*. No new identifications are as yet made from the 1958 collections of arctolepids from Ekmanfjorden.

Recognisable cephalaspid fragments have not been found, though Wängsjö (1952) claims *Nectaspis* as a guide fossil to the Lykta Division and described material from the area.

The Age of the Kapp Kjeldsen and Lykta Divisions

At Sigurdfjellet in Woodfjorden the Kapp Kjeldsen Division rests apparently conformably upon the highest part (Dittonian) of the Red Bay Series and is dated as lower Lower Devonian by Føyn and Heintz (1943, p. 47). The Dittonian type area in Britain (White, 1950, 1956) has been subdivided into pteraspid zones, none of which is directly recognisable in Spitsbergen though there seem to be similarities in the general sequence of faunas in the two regions. Pteraspids in Britain range from the basal Dittonian beds to the Senni Beds, in Spitsbergen from mid Fraenkelryggen Division to Wijde Bay Series (Dineley, 1956). Further discoveries of pteraspids in the Wijde Bay Series have recently been made by a Cambridge expedition (P. F. Friend, personal communication).

Giganthaspis, the typical pteraspid of the Kapp Kjeldsen Division, may be related to *Pteraspis* (*Rhinopteraspis*) *dunensis* but its affinity with "*P. cornubicus*" is by no means so easily determined. (Dr. E. I.

White's current studies of the latter species throw doubt upon its taxonomic status.) "*P. cornubicus*" is from beds regarded as upper Gedinnian to middle Siegenian in age (upper Dittonian) while *P.* (*R.*) *dunenis* (*s.s.*) is middle Siegenian and younger (Breconian.) There may then be grounds for suggesting that the *Giganthaspis*-yielding beds are of lower or middle Siegenian age (uppermost Dittonian—Breconian).

While *Doryaspis* appears as an excellent guide-fossil to the Lykta Division, it gives no direct indication of the age of the group, being unknown outside Spitsbergen. On general stratigraphical grounds, therefore, a Breconian (m. Siegenian-Emsian) age for the *Doryaspis* beds may be suspected.

Structure

The Old Red Sandstone along the eastern side of Ekmanfjorden is thrown into broad N.-S.—striking folds. Dips seldom reach 25 degrees; usually they are a little less than 15. Since they often have a strong easterly component they appear from the fjord to be inclined at smaller angles. The northernmost flexure exposes the Kapp Kjeldsen Division on the lower western flanks of Orsafjellet and Rattvikfjellet. Here the dips are eastwards at up to 25 degrees, swinging round to south-east at the south-western end of Rattvikfjellet and probably to north at the north-western end of Orsafjellet. On the mountain sides south-west of Gygrestolen a gentle synclinal arrangement of the beds is observed but again the strata are also tilted at about 10 degrees towards the east. At Garborgnuten the higher beds of the Lykta Division dip at about 10 degrees east-north-eastwards. South of here the strata in the mountain sides are tilted eastwards at between 15 and 25 degrees. On the coastline the dip is from 10 to 15 degrees. Local normal faults occur here and there and downthrow only a few feet. The folding of the strata is clearly pre-Carboniferous. Its northerly strike suggests a possible connection with the (Upper Devonian) boundary faults of the central Spitsbergen N.-S. graben walls (see insert text-fig 1). Large boulders of Hecla Hoek rocks present in the moraine of Orsabreen may indicate that the pre-Devonian basement perhaps outcrops within the field of this glacier. Locally, however, there is no indication that the basement rocks are involved in the folding.

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Notes on the distribution of English Bathonian Foraminifera

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Abstract

The English Bathonian contains a rich foraminiferal fauna composed largely of Lagenidae. The succession of foraminifera is as yet only partially revealed, but some faunal changes between the Lias and Bathonian are apparent, and four faunules have been distinguished in the Bathonian. In Dorset and the vicinity of Bath the faunules occur in the same order of superposition. In comparing the English Bathonian foraminifera with those in the German section the present data suggest that a number of species range higher in England than in Germany.

Introduction

The purpose of this paper is to summarize the distribution of English Bathonian foraminifera. It is based on collections of foraminifera made by the writer from outcrops extending from the Dorset Coast to Rutland. Descriptions of species and records of their stratigraphic occurrences are presented elsewhere (Cifelli, in press).

The strata of the Bathonian are generally poorly exposed and part of the sequence is completely concealed. The distributional data, therefore, are incomplete and many additional records of species are needed to establish the foraminiferal succession in this stage. Nevertheless, the present records reveal some important aspects of the Bathonian fauna and should provide a basis of comparison for future investigations.

General Aspects

In summarizing the distribution of Bathonian ammonites Arkell (1951, p. 15) has stated : " Thus in Britain as the world over the Bathonian shelf seas were for the greater part inimical to ammonites. It was a period of restriction in ammonite distribution when the waters were invaded by deltas of great rivers, an environment where cephalopods never thrived, and in the deposits of which they are generally absent or rare." The conditions which restricted the distribution of ammonites do not appear to have affected the foraminifera adversely, as the Bathonian deposits of England contain a rich and varied foraminiferal fauna, not differing fundamentally from that of the Lias.

Among the Bathonian foraminifera, the Lagenidae are by far the most important, outnumbering all other families in individuals, species and genera. In numbers of individuals the Lagenidae are rivalled only by coiled, calcareous tubular forms belonging to either *Cornuspira*

liasina Terquem or *Spirillina infima* (Strickland). Occasional samples of foraminifera were observed to contain as many specimens of these two species as of all the Lagenidae combined. The Polymorphinidae constitute an important subordinate element of the Bathonian fauna, but are of simple structure, and all of them are referable to the primitive genera *Eoguttulina* and *Paleopolymorphina*. A few calcareous imperforate species belonging to structurally simple genera of Ophthalmidiidae and Miliolidae are locally common. Arenaceous forms seldom occur in appreciable numbers, and are represented by few genera and species. The Rotaliidae exhibit little diversification, and individuals are not common, except for occasional floods of *Epistomina stelligera* (Reuss) found to occur at several horizons. The only Buliminidae observed in the English Bathonian are rare specimens of *Bolivina rhumbleri* Franke.

Lateral changes in facies are exhibited by the Bathonian foraminiferal assemblages which correspond in a general way to the lithological changes occurring in the Great Oolite Series. North of Bath, where the Bathonian is composed mostly of the Great Oolite Limestones, foraminifera are less numerous and are represented by fewer genera and species than in the dominantly clay sections of the Bath area and the Dorset Coast. Many diagnostic forms of the southern areas have not been observed in the Cotswolds, Oxfordshire or Northamptonshire. The ornamented species in particular are seemingly more common in the southern areas. On the other hand, species in the Great Oolite Limestone appear to be, on the whole, larger and more robust. Particularly common in the northern areas is *Vaginulina legumen* (Linné), which diminishes in abundance towards the south and is rare in the Bathonian of the Dorset Coast.

Despite facies changes the assemblages are dominated in practically all instances by the Lagenidae. The one observed exception occurs in the Upper Estuarine Beds of Rutland, where the dominant forms are *Eoguttulina liassica* and *Spirillina infima*. In the upper part of the Upper Estuarine Beds at the Clipsham New Quarry, Rutland, the assemblage is composed entirely of *Eoguttlina liassica*.

The dominance of the Lagenidae in the Bathonian is characteristic of the Jurassic foraminiferal faunas, which contrast with modern ones in which the Lagenidae are of only secondary importance. It has been suggested (Glaessner, 1945, p. 190 ; Barnard, 1950, p. 40) that the Lagenidae occupied a wider ecologic range in the past than at the present. The modern Lagenidae, however, are a family of widespread and frequent occurrence. They are probably most varied and abundant in temperate waters of the continental shelf between depths of 50 and 500 fathoms, but are common at 5 to 60 fathoms and persist in reduced numbers to great depths (Kleinpell, 1938, pp. 14–15; Glaessner, 1945,

p. 187; Cushman, 1948, p. 44). Individual species and genera are, of course, much more restricted in their distribution.

That the Bathonian Lagenidae as a family had a tolerance for a large variety of environments is indicated by the variety of sediments in which they occur. There are no indications, however, of extreme ranges in temperature or depth in the English Bathonian, and the Great Oolite was probably deposited in tropical to subtropical waters of shallow depths (Woodward, 1894, pp. 16–21). It is doubtful if the depths ever exceeded those of our present continental shelves.

The dominance of the Lagenidae in practically all marine environments represented in the English Bathonian, as in the remainder of the Jurassic throughout the world, is probably best accounted for by the lack of competition from other foraminifera. It was a time of minimum selection pressure for the Lagenidae when this euryopic family was able to exploit a wide variety of habitats in which they rapidly multiplied and exhibited much variation. Their dominance continued through the Lower Cretaceous to the Aptian, when a large number of Buliminidae and Discorbidae appear (Glaessner, 1945, p. 204). Since the Lower Cretaceous the importance of the Lagenidae has decreased, and other families, particularly the Rotaliidae and the Buliminidae, have assumed important roles in the foraminiferal assemblages. The modern Lagenidae, though widely distributed, constitute but accessory elements in most of the assemblages in which they occur.

Important evolutionary changes among the calcareous families which later surpassed the Lagenidae probably occurred in the Jurassic as well as in the Lower Cretaceous. Their early history, however, is incompletely known and important evolutionary changes may have occurred in the deeper waters below the continental shelves. Deposits of these environments are rare or absent in the Jurassic in most of Europe, but may be present in the geosynclinal areas of the Alps.

Comparison with the English Lias

Most of the foraminiferal faunas of the English Jurassic are yet to be described, so that it is not possible at present to determine the total ranges of the Bathonian species in England, or to make detailed comparisons of the Bathonian fauna with those of other stages.

Listed below are characteristic species occurring in the English Bathonian, but which have not been previously recorded from the English Lias. The list is not complete, but includes only those species that are reasonably common and which are believed to be sufficiently diagnostic as not to be confused with others occurring in the Lias.

Citharina heteropleura Terquem.
Dentalina intorta Terquem.

Dentalina oolithica Terquem.
Epistomina stelligera (Reuss).
Flabellammina althoffi Bartenstein.
Lenticulina quenstedti (Gümbel).
Lenticulina tricarinella (Reuss).
Massilina n. sp.
Nodosaria opalini Bartenstein.
Patellina oolithica Terquem.
Pseudonodosaria hybrida (Terquem and Berthelin).
Spiroloculina lanceolata (Terquem and Berthelin).
Spirophthalmidium concentricum (Terquem and Berthelin).
Triplasia bartensteini (Loeblich and Tappan).
Trochammina haeusleri (Galloway).
Vaginulina clathrata (Terquem) n. subsp.

Since the Bajocian foraminiferal fauna has not as yet been described, the lower ranges of these species are not known; some of them may occur in the Inferior Oolite. The ranges of these species above the Bathonian are similarly poorly known. There are a few descriptions and faunal lists of the younger Jurassic foraminifera in the older literature, but these are not adequate for comparison. Recently Barnard (1952 and 1953) described the foraminifera from the Upper Oxford Clay at Warboys, Huntingdonshire and Weymouth, Dorsetshire. *Lenticulina quenstedti* is the only species from the list above which occurs in the Upper Oxford Clay at Warboys ; none were recorded at Weymouth.

Among the Lias foraminifera the following characteristic species have not been observed in the English Bathonian:—

Dentalina matutina d'Orbigny.
Frondicularia bicostata (d'Orbigny).
Frondicularia brizaeformis Borneman.
Frondicularia sulcata Borneman.
Lingulina tenera Borneman (sensu lato).
Marginulina prima d'Orbigny.
Nodosaria oculini (Terquem and Berthelin).
Ophthalmidium northamptonensis Wood and Barnard.

Perhaps the most important one of these which is absent in the Bathonian is *Lingulina tenera* and related forms which Barnard (1956, p. 275) has called the *Lingulina tenera* " plexus ". The genus *Lingulina* is not common in the Bathonian and the species are mostly smooth, with little or no ornamentation.

Several species characteristic of the Lias appear only in the Lower

Bathonian, and have not been observed above the lower part of the Lower Fuller's Earth Clay. These are :—

Nodosaria liassica (Barnard).
Frondicularia lignaria (Terquem).
Vaginulina clathrata (Terquem) sensu stricto.

The lists above summarize some of the more important ways in which Bathonian faunas may be distinguished from those of the Lias. There are no fundamental differences between the Bathonian and Lias faunas. *Lingulina* and *Frondicularia* appear to have attained a more complex development in the Lias, where they are more common and the species exhibit more complex ornamentation. *Lenticulina*, on the other hand, is better developed in the Bathonian where there are more species, including the ornamented forms *Lenticulina quenstedti* and *L. tricarinella*.

Summary of the Foraminiferal Sequence

Portions of the Bathonian are not represented because of the lack of exposures, and many details of the faunal sequence are lacking. Consequently, a formal subdivision of the strata based on foraminifera is hardly justified at the present. Nevertheless, an analysis of the known stratigraphic occurrences of species reveals that in the areas of Bath and the Dorset Coast there are four recognizable faunules which in both areas succeed each other in the same order. North of Bath the assemblages appear much less distinctive and these faunules cannot, on the known records of species, be recognized. The northern assemblages differ from those to the south primarily in negative characters. That is, they are less rich and are generally represented by fewer species and individuals. The preservation of the northern samples, coming as they do mostly from marls and calcareous clays, is generally poor and perhaps some of the faunal distinctions may be more apparent than real.

There are still but few data on the stratigraphic occurrences of the species of these faunules in most parts of Europe. Consequently, the ranges of the species are incompletely known, and the relationships of the faunules to larger faunal units (based on foraminifera) existing in the European Province cannot as yet be determined.

The tables showing the stratigraphic occurrences and taxonomic details of the species of the faunules are given elsewhere (Cifelli, in press). The general aspects of the faunules and their stratigraphic relationships are summarized below. They are arranged in ascending order.

Faunule A.—Species not recorded from older strata in England, and apparently making their first appearance in this faunule, are: *Dentalina*

oolithica, Lenticulina quenstedti, L. tricarinella sensu lato, *Nodosaria opalini* and *Trochammina haeusleri.* Species characteristic of the Lias which occur in, but have not been observed to extend beyond this faunule, are: *Vaginulina clathrata, Frondicularia lignaria* and *Nodosaria liassica.* The following forms have not been observed or recorded beyond the limits of this faunule in England and appear to be restricted to it: *Vaginula macilenta, Frondicularia nodosaria* n. var. and *Lenticulina quenstedti* var. A. (n. var.). Although not restricted, the following species are common here: *Lenticulina galeata, Nodosaria hortensis, Spirophthalmidium concentricum.*

This faunule occurs in the basal few feet of the Lower Fuller's Earth Clay. It was observed at Burton Bradstock in Dorset and in the railway embankment at Midford, near Bath. *Oppelia fallax* and related ammonites occur in the basal beds of the Lower Fuller's Earth, but Arkell (1951, p. 20) does not consider them diagnostic and he has assigned these beds to the *Zigzag* zone. Most of the Fuller's Earth Clay is never exposed and consequently there is an important gap in the foraminifera and ammonite sequences. No foraminifera have as yet been recovered from the Lower Fuller's Earth Clay north of Bath.

Faunule B.—Species from the previous faunule which have not been observed to range beyond this one are : *Citharina colliezi, Nodosaria* aff. *N. prima, Lenticulina quenstedti* var. B. (n. var.) and *Planularia eugenii.* Appearing for the first time and not recorded from the prior faunule are *Dentalina propinqua, Nodosaria pectinata, Frondicularia spissa,* and *Vaginulina clathrata* n. subsp. Occurring commonly, though not restricted to this faunule, are: *Planularia beierana, Lenticulina tricarinella,* and *L. munsteri.*

This faunule has been recognized in the upper beds of the Lower Fuller's Earth Clay, the Fuller's Earth Rock and the *Wattonensis* Beds. At Dorset it occurs in the *Wattonensis* Beds at the foreshore of Watton Cliff. It is well developed in the upper beds of the Lower Fuller's Earth Clay and in the Fuller's Earth Rock in the Maperton–Charleton–Horetorne road-cut in Somerset. It was also observed in the town of Box and at the Cross Ways Inn road intersection, about 5 miles south of Bath. Consequently, it occurs both in, and immediately below, strata belonging to the *Subcontractus* Zone. Since no ammonites have been found in the upper part of the Lower Fuller's Earth Clay, the relationships of these beds are not clear.

Faunule C.—Species occurring in this faunule which were not observed in the previous ones are *Citharina heteropleura* and *Dentalina* aff. *D. communis.* Species which occur in this and earlier faunules but not in the succeeding one are: *Vaginulina harpa, Citharina* sp., and *Nodosaria pectinata.* Species that are common in this faunule are:

Dentalina intorta, Nodosaria opalini, Epistomina stelligera and *Lenticulina subalata.*

This faunule was observed in the Upper Fuller's Earth Clay. In Dorset it occurs at Watton Cliff and Cliff's End. In the vicinity of Bath it was observed in the Combe Hay Fuller's Earth Works, Vernham Wood, and Henley Hill, near Box. In the Bath area the Upper Fuller's Earth Clay represents the upper part of the *Subcontractus* Zone and is overlain by the Great Oolite Limestones of the *Aspidoides* Zone (Arkell, 1951, p. 22). In Dorset, however, there are no Great Oolite Limestones and the Upper Fuller's Earth Clay is overlain directly by the Forest Marble. There may be an unconformity between these two formations in the Dorset area representing a hiatus equivalent in time to the *Aspidoides* secule. It is also possible, however, that the Upper Fuller's Earth Clay replaces the Great Oolite Limestones in the Dorset area. In this case the top of the Upper Fuller's Earth Clay would be younger in Dorset than it is in the Bath area. Amonites do not occur in the Upper Fuller's Earth Clay of the Dorset Coast. In view of these uncertainties all that can be said concerning the relations of faunule *C.* is that it occurs in the *Subcontractus* Zone and in strata of uncertain age which may belong to the upper part of the *Subcontractus* Zone, or younger strata, possibly belonging to the *Aspidoides* Zone.

Faunule D.—This faunule is similar to the preceding. The most characteristic feature is the common and apparently restricted occurrence of *Massilina* n. sp. This species was not observed in any other part of the English Bathonian and there are no records of it in any other strata. *Nodosaria clavula* and *Frondicularia intumescens* have not been observed elsewhere in the Bathonian. These species are rare, however, and *Frondicularia intumescens* occurs in the Lias. *Vaginulina clathrata* n. subsp. and *Dentalina subplana* make their last appearance in the *Boueti* Bed and the Bradford Clay, the basal strata in which this faunule occurs.

This faunule characterizes the Forest Marble, including the *Boueti* Bed and the Bradford Clay. In Dorset it occurs in the Forest Marble section exposed at Watton Cliff and in the *Boueti* Bed at Herbury, along the West Fleet. The ammonite *Clydoniceras hollandi*, index species for the *Hollandi* Zone, occurs in the Bradford Clay (Arkell, 1951, p. 22) and therefore, faunule *D* occurs in this zone. Traditionally, the *Boueti* Bed of Dorset has been correlated with the Bradford Clay and thus placed in the *Hollandi* Zone. However, Sylvester-Bradley (1957, p. 27) has correlated the Bradford Clay with the *Digona* Beds of Dorset, a horizon higher in the Forest Marble and he has placed the *Boueti* Bed in the *Aspidoides* Zone, equivalent to the Great Oolite Limestones of the Bath area. Since no foraminifera were recovered from the Great Oolite Limestones they cannot shed much additional

light on this problem. Faunule *D* makes its first appearance in the *Boueti* Bed of Dorset and in the Bradford Clay in the Bath area. In Dorset faunule *D* immediately overlies faunule *C* which occurs in the Upper Fuller's Earth Clay. In the Bath area however, faunule *D* is separated from faunule *C* occurring in Upper Fuller's Earth Clay by the Great Oolite Limestones, and the stratal limits of the two faunules are unknown in this area. About all that can be said with reasonable certainity on the basis of foraminiferal evidence is that the *Boueti* Bed of Dorset is younger than the Upper Fuller's Earth Clay of Bath.

Comparison with the Dogger of North-west Germany

The most comprehensive stratigraphical analysis of Jurassic foraminifera is of the Lias and Dogger of North-west Germany, where Bartenstein and Brand (1937) described and recorded the stratigraphic ranges of a large number of species. Many of the species occurring in the Dogger of North-west Germany also occur in the Bathonian of England, and some faunal comparisons between the two regions are possible. In this discussion the taxonomy is that of Cifelli (in press) and involves some changes to species names used by Bartenstein and Brand.

In Germany, the Dogger is a lithogenetic unit comprizing all of the Middle Jurassic, and includes strata ranging in age from Bajocian through Callovian. Only the *Wurttembergicus* Schichten and the *Aspidoides* Schichten, the middle lithologic divisions of Dogger Epsilon, are Bathonian.

Bartenstein and Brand (1937, p. 114) divided the Dogger into four faunal divisions based on foraminifera, which, however, they did not formally designate as zones. The boundary between two of these divisions (Abteilung III and IV) occurs between the *Wurttembergicus* and *Aspidoides* Beds. According to Arkell (1956, pp. 140–141) the *Wurttembergicus* Beds represent the *Zigzag* Zone (lowermost Bathonian) while the *Aspidoides* Beds range in age from Middle through Upper Bathonian, probably including beds representing the *Subcontractus* Zone in the lower part. In England roughly comparable strata can be distinguished on the basis of foraminifera (faunules A and B). The criteria for distinction, however, are not the same in the two regions. Although there are many species in common between England and Germany, the ranges of certain species appear to differ somewhat in the regions.

The fauna of the lowermost Bathonian (*Wurttembergicus* Beds) in Germany is closely allied to that of the Bajocian, and Abteilung III of Bartenstein and Brand extends across the boundary of the two stages. Many species characteristic of the English Bathonian make their first appearance in Germany in the Bajocian. Among these are *Lenticulina*

quenstedti which first occurs in Dogger Gamma and *L. tricarinella* which is first recorded from the *Parkinsoni* Schichten of Dogger Epsilon. In England, the earliest records of these species are from the Lower Bathonian (basal beds of the Lower Fuller's Earth Clay). Possibly they will be discovered in the Inferior Oolite, when the Bajocian faunas ofEngland are described.

Of particular interest are the apparent " holdovers " that appear to have persisted in England during the Bathonian. If ammonite correlations are used as a basis of control a number of species appear to range higher in England than they do in Germany. The differences in the ranges of the important species are summarized below, but no reason for the discrepancies in ranges can be given at the present time.

Vaginulina macilenta is restricted in North-west Germany to the Bajocian (Dogger Alpha through Delta), but in England it occurs in lowermost Bathonian (basal beds of the Lower Fuller's Earth Clay).

Vaginulina clathrata is succeeded by *V. clathrata* n. subsp. in England and Germany. The succession of these closely related forms which overlap morphologically is highly suggestive in each region of a chronologic subspecies. However, the succession occurs earlier in Germany than it does in England. *Vaginulina clathrata*, a common form in the Upper Lias, appears to have its highest occurrence in the Lias Zeta of North-west Germany, but ranges into the *Zigzag* Zone of the Lower Bathonian in England. *V. clathrata* n. subsp. seemingly first occurs in the Lower Bajocian (Dogger Alpha), while in England it is first recorded in the Middle Bathonian (Fuller's Earth Rock).

Nodosaria opalini is restricted to the Lower Bajocian (Dogger Alpha and Beta) in Germany. In England this species occurs throughout the Bathonian and is particularly common in the Middle Bathonian (Upper Fuller's Earth Clay).

Discorbis dreheri ranges only from Lias Zeta through Dogger Beta in Germany, but occurs throughout the English Bathonian.

Vaginulina hechti has been recorded by Bartenstein and Brand from the Dogger Alpha and Beta. One fragment of this species was observed in the *Wattonensis* Beds (Middle Bathonian).

Anomalina liassica which ranges from Lias Zeta to Dogger Beta in Germany occurs (though rarely) in the Upper Fuller's Earth Clay.

Bolivina liassica and *Frondicularia intumescens*, species characteristic of the Lias, and restricted to this interval in Germany both occur in the Bathonian of England. *Bolivina liassica* was observed in the *Wattonensis* Beds, and *Frondicularia intumescens* occurs in the Forest Marble (Upper Bathonian).

Nodosaria issleri is restricted to the Lias in Germany. In England, forms differing but slightly, though perhaps not identical (*Nodosaria*

aff. *N. issleri*), occur in the Upper Fuller's Earth Clay (Middle Bathonian).

A few forms appear to range higher in Germany than in England. *Vaginulina clathrata* n. subsp. was not observed in strata younger than the *Boueti* Bed (lower Upper Bathonian), while in Germany identical forms occur in the Dogger Zeta (Lower Callovian). *Citharina heteropleura* also occurs in the Dogger Zeta, while it has not been observed in England above the Upper Fuller's Earth Clay. *Lingulina pupa* was not seen in the English Bathonian, but Bartenstein and Brand report rare occurrences of this species throughout the entire Dogger.

There are two characteristic species which are not common to the two regions. *Epistomina mosquensis* is common throughout most of the German Dogger, but was not observed in the English Bathonian, and has never been recorded from any English strata. *Massilina* n. sp., a common form in the Upper Bathonian of England, has not been recorded from Germany.

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Cornstone

By J. R. L. Allen

Abstract

Two dissimilar rocks have been confused under the title "cornstone": one is a concretionary limestone, the other a calcareous conglomerate. A return to the names "concretionary cornstone" and "conglomeratic cornstone" is advocated since the rocks differ in their significance. Conglomeratic cornstones, intraformational in character, point to sudden increases in current activity. Concretionary cornstones suggest declines in the supplies of clastic detritus.

Introduction

"Cornstone" is one of the many rock names taken from the English vernacular. The term, once widespread in the country about the River Severn and Bristol Channel, was adopted for scientific use early in the nineteenth century. Originally the word probably meant a granular rock (Arkell and Tomkeieff, 1953, p. 29), though a less likely meaning akin to that of "cornbrash" has been suggested (Page, 1865, p. 154; North, 1930, p. 329). At present geologists use "cornstone" for two distinct kinds of rock.

Definition and Use

Greenough (1819, p. 218) probably made the earliest use of "cornstone" in a scientific sense, but Buckland (1821, p. 512) was the first to define it clearly. Buckland's cornstone :—

> "is composed of marl or marlstone, filled with concretions of compact limestone, presenting the fracture and colour of mountain limestone, and varying in size from that of a pea to blocks of many tons, and sometimes spreading itself into thick and compact beds, to the almost total exclusion of the marl. The knotted character which these concretions assume resembles that of a conglomerate animal gland, and the small acini or kernels of which they are composed usually separate under the blow of the hammer. The transfusions of their outer portions and projecting points into the substance of the marlstone, shows them not to be fragments resulting from the destruction of any older rocks of transition limestone, but concretions of contemporaneous origin with the marlstone, in which they are imbedded."

The rock Buckland referred to cropped out in the Lickey Hills near Birmingham and was given by him an Old Red Sandstone age. Murchison (1839, p. 55) correctly assigned the bed to some part of the New Red Sandstone, identifiable either as the Halesowen or Keele Group. Since Buckland's clear statement, calcareous deposits in the

Old Red Sandstone and Permo-Carboniferous of Great Britain have been called " cornstone ". Murchison confused matters by applying the name to calcareous conglomerates. Unfortunately, he used the term in Buckland's sense as well, describing concretionary and conglomeratic rocks in the same strata under the one head. Most of his contemporaries, for example Malcolmson (1838, p. 668) and Phillips (1848, p. 131), adhered to Buckland's definition.

M'Cullough (1869, p. 8) realized that in the Old Red Sandstone two dissimilar rocks had been confused. He therefore distinguished a " concretionary cornstone " (calcareous concretions embedded in marls and grading to solid concretionary limestones) and a " conglomeratic cornstone " (marl and limestone fragments embedded in more or less sandy and calcareous matrices). The Geological Survey also used these names in the same sense for concretionary and conglomeratic rocks in the Old Red Sandstone and Permo-Carboniferous (Strahan and others, 1909, pp. 72–73 ; Dixon, 1921, pp. 31–34 ; Barrow and others, 1919, pp. 53–54).

Most workers neglected M'Cullough's proposals and continued to use the ambiguous " cornstone ". Sollas (1879, p. 491), Goodchild (1904, p. 118), Greenly (1919, p. 579), Wallis (1927, p. 727), and Capewell (1957, p. 399) applied it correctly in Buckland's sense to the concretionary rocks in the Old Red Sandstone. King (1925, p. 386 ; 1934, p. 528), Trotter (1942, p. 9), Clarke (1951, p. 101), Dineley (1951, p. 140), and Ball and Dineley (1952, p. 210) followed Murchison, using " cornstone " for the conglomeratic deposits. Other workers have written of both rocks under the one head.

Discussion

It is clearly undesirable that two such dissimilar rocks should be known by one name. Both types of deposit are key constituents of the Old Red Sandstone and Permo-Carboniferous. The writer's studies of Dittonian and Breconian strata in the Old Red Sandstone of South Wales and the Welsh Borderland show, for example, that conglomeratic cornstones normally lie at the bases of thick cyclothems (conglomerate → sandstone → siltstone) and therefore denote the renewal of cycles of sedimentation. Concretionary cornstones, on the other hand, often lie amongst the higher beds of the cyclothems, suggesting that the supply of clastic detritus waned periodically.

Usage and priority urge a return to Buckland's definition of " cornstone ". Against this is its widely current application in Murchison's sense. The best solution seems to be M'Cullough's, relinquishing the use of the unqualified " cornstone ".

To promote this, general re-descriptions of the rocks are given below, placing emphasis on Old Red Sandstone examples.

Concretionary Cornstones

These, the original cornstones of Buckland (1821), are abundant in the Old Red Sandstone of the Midlands, the Borderland, Wales, Scotland, and Ireland, and are also found as some of the *Spirorbis*-bearing limestones in the red-bed measures of the coalfields of the Midlands. The rocks are mottled red, green, or blue and form wedging units averaging about 5 ft. thick, but reaching locally about 30 ft. as in the Red Marl Group at Llanstephan, Carmarthenshire. Fossils are absent in the deposits from the Old Red Sandstone. Most of the units, as in Corvedale, Shropshire, and the Golden Valley, Herefordshire, commence with small, knotted, irregular concretions of argillaceous limestone embedded in massive siltstone. These concretions increase in size upwards, usually coalescing finally to solid or rubbly limestones bearing pods or patches of siltstone. The limestones are unlaminated and, like the concretions, often show "cats-brain" mottling. Strong vertical joints are typical. Most of the concretionary cornstones, both in the Old Red Sandstone and the Permo-Carboniferous, are underlain by siltstones. Siltstones normally follow and grade up rapidly from the top limestones. Sandstones rest occasionally on the top beds, an erosion surface often lying between. The concretionary cornstones may be said to grade from concretionary calcareous siltstones to concretionary argillaceous limestones.

Under the microscope the concretions and their enveloping siltstones, typified by rocks from near Holdgate, Corvedale, appear unlaminated and structureless. Rare to abundant quartz of silt to fine sand grade and a little muscovite or biotite are scattered unevenly in the concretions. They are accompanied by brown and green clay minerals, sometimes in multicrystalline aggregates. Dusty to microcrystalline calcite forms the groundmass. Crystals of dolomite are rare. Greenly (1919, p. 580), King (1925, p. 385), and Heard and Davies (1924, p. 505) found by weight analysis of typical rocks from 2–91 per cent of calcite and from 1–7 per cent of dolomite. Sharp veins of calcite, often zoned by drusy crystals, cut the concretions and branch into veinlets ending chiefly in the dusty calcite. Quartz grains, embedded in a dominant groundmass of brown and green clay, are evenly scattered in the siltstones. Irregularly distributed patches of calcite are sometimes present. As they increase in abundance and size the siltstones grade slowly into the concretions.

The precise mode of origin of the concretionary rocks is uncertain though they were obviously laid down under quiet conditions. They may have been formed by the simultaneous mechanical deposition of silt, clay, and precipitated carbonates in evaporating bodies of water, cut off from major supplies of clastic detritus. Suggestively, siltstones associated with concretionary cornstones in the Old Red Sandstone

often bear sun-cracks. Against this is the unlaminated character of the limestones, though of course some post-depositional process may have destroyed any original stratification. Alternatively, the deposits might be cemented muds, originally " soupy " at the sediment-water interface but increasingly more compacted in depth. In this case the amount of the carbonates would roughly represent the porosity of the deposits at the time of cementation.

Conglomeratic Cornstones

Conglomeratic cornstones, the " cornstones " of Murchison (1839), are also widespread in the Old Red Sandstone of the Midlands, the Borderland, and Wales, and in the Permo-Carboniferous red-beds of the Midlands. They are rare in the Old Red Sandstone of Scotland and Ireland. Both in the Old Red Sandstone, as at Lligwy Bay, Anglesey, and in the Permo-Carboniferous, as in the Clent Hills and at Hunnington, near Birmingham, the rocks form tabular or wedging units averaging about 1 ft. in thickness. Units of 20 ft. occur in the Old Red Sandstone of the Forest of Dean. Typically the beds grade up into sandstones. They commonly rest on siltstones, erosion surfaces separating the two. Some units, especially in the Old Red Sandstone, are cross stratified whilst others reveal an imbrication of the fragments and bear lenses of sandstone, occasionally ripple-marked.

Normally the dominant fragments are of red, green, or brown siltstones. Those of white, pink, or yellow argillaceous limestones are usually less common, but in some units, chiefly those in the Old Red Sandstone, exceed the siltstones in amount. The deposits, therefore, vary from siltstone-limestone to limestone-siltstone conglomerates. Fragments of sandstone, vein-quartz, and exotic rocks are usually absent and never become abundant. Conglomeratic cornstones in the Old Red Sandstone often carry broken or worn " fish " scales and plates. In size the fragments range from granules to boulders. They are well sorted, the majority in each unit falling within a single class (granule, or pebble, or cobble). Boulders are infrequent and always of siltstone. The particles are angular to well rounded and the largest are best rounded. At the same size, the fragments of siltstone are normally better rounded than those of limestone. Typically quartz sand or silt fills the voids between the fragments whilst coarsely crystalline calcite is usually an abundant cement. Some conglomeratic cornstones in the Old Red Sandstone bear only calcite in the gravel open-work.

Most of the gravel materials are petrographically indistinguishable from the siltstones and concretionary limestones bedded in the same successions. The eroded surfaces beneath many of the conglomeratic cornstones prove that denudation of the limestones and siltstones took place. The conglomeratic cornstones are therefore intraformational

conglomerates (Walcott, 1894, p. 191) and point to the renewed action of strong currents in a depositional basin, rarely concomitant with floods of coarse detritus from distant sources. Those of the Old Red Sandstone compare well, for example, with vertebrate-bearing limestone-mudstone conglomerates in the continental Moenkopi Formation (Triassic) of Arizona (McKee, 1954, p. 45).

Conclusion

The unambiguous names " concretionary cornstone " and " conglomeratic cornstone " were widely used at an earlier date. Their revival would remove the present confusion, permitting a clear distinction between two rocks of different origin.

Many friends and colleagues gave the writer their encouragement whilst this note was in preparation. Thanks go to these, and especially to Dr. L. B. Tarlo and Dr. R. Goldring.

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A Formation of Marekanite at El Fiscal, Guatemala

By Pontus Ljunggren

(PLATE I)

Abstract

An outcrop of obsidian pebbles in a grey volcanic rock was investigated in the field and subsequently examined in the laboratory by means of optical, X-ray, and D.T.A. methods. The results show that the grey rock arose from obsidian through a micro-cracking, causing a formation of marekanite balls (massive obsidian) surrounded by a very brittle grey rock (micro-cracked but completely unaltered volcanic glass).

Introduction

In the vicinity of El Fiscal, some 25 km. from Guatemala City, along the Atlantic Highway to Puerto Barrios, there are some large outcrops of volcanic rocks dating from the early Quaternary period (cf. Sapper, 1937) and including glassy varieties of obsidian type. The dark and glassy obsidian occurs as rounded " pebbles " and " stones " in a grey, volcanic rock and not as homogeneous layers. It is, moreover, concentrated in certain layers within this grey rock (Plate I, Fig. 1). The grey rock is extremely brittle and often has a spherulitic texture. It readily splinters into needles up to one inch long. When struck by the hammer the grey rock turns into fine powder.

Laboratory examinations of the two rocks have been carried out at the Geological Institute of the University of Lund, and the results are given in the present paper.

Obsidian

The obsidian of El Fiscal is characterized by a varying frequency of microlitic minerals in the volcanic glass. These microlitic minerals are dominated by feldspars together with smaller amounts of biotite, the grain-sizes varying from 0·1 to 0·03 mm. or less. The feldspars are untwinned and quite unaltered; moreover, they often show a skeletal development. A refractive index determination and X-ray examination revealed the characteristics of albite. All the feldspars are euhedral. The pleochroism of the biotite is γ dark brown and α yellowish brown. There are no inclusions in the albites nor in the biotites.

Other minerals of a smaller grain size than the microlites also exist in the volcanic glass. A few of these crystallites can be identified as magnetite (the quite predominating opaque mineral), grain size 0·01–0·001 mm.

The frequency of microlites and crystallites varies greatly in different specimens, from almost pure glass to a glass with about 50 per cent

crystallites. These inclusions are arranged in parallel bands representing the flow direction of the lava. The sheen, or the reflection of light from within the obsidian specimens when held in certain positions, is due to this parallel arrangement of the albite microlites.

X-ray powder photograms of the obsidian of El Fiscal gave, besides the weak pattern of albite, a broad band at about 4·0Å, and, in some specimens which are almost free from microlites and crystallites, one additional band at about 7·4Å.

The density of the obsidian is 2·37, giving a content of SiO_2 of 68 per cent (W. O. George, 1924). The refractive index is $n_{Na} = 1·489$, corresponding to a SiO_2 content of 75 per cent (*op. cit.*).

The loss upon ignition is 6·0 per cent.

The Grey Rock

A microscopic examination of the grey rock revealed that it has exactly the same mineralogical and textural composition as the obsidian. An X-ray examination gave the pattern of albite and one or two diffuse bands representing the glassy compound. A D.T.A. analysis gave no deflections up to 1,000° C.

The density of the grey rock is 2·35, thus 0·02 less than that of the obsidian. This indicates that, if the two rocks are of identical composition, the grey rock has a porosity of about 1 per cent. When a fragment of the grey rock (D = 2·35) is immersed in a heavy liquid (acetylene tetrabromide and benzene, D = 2·37), it first floats. After five minutes or so, however, it has attained the density of the liquid (and of the obsidian) through adsorption of liquid into the pores.

A microphotograph covering both sides of the boundary grey rock—obsidian is given in Plate I, Fig. 2.

The loss upon ignition is 7·5 per cent.

Discussion

The examinations reported above give clear evidence that the grey rock is a porous variety of the obsidian, but it was also found that there is no difference in the composition of the glasses nor in the mineralogical composition of the respective rocks. The grey rock is extremely brittle, this being due to a large number of sub-microscopic cracks. It is evident that this rock arose through a cracking of the original obsidian glass. This cracking was not complete, however, and rounded " pebbles " and " stones " of obsidian were left within the former obsidian lava flow.

Microscopic examination has shown that the glass of the present obsidian is completely isotropic, and heating-cooling experiments have further supported the opinion that this glass is unaffected by strain. However, the characteristics of the grey rock indicate that this

rock arose from obsidian through a release of strain, leaving only strain-free nodules: the present obsidian " pebbles " and " stones ".

The plagioclase microlites of both rocks, as also most of the biotites, are arranged parallel to the flow faces of the original lava, indicating that these minerals crystallized before the final solidification of the obsidian lava. There is no evidence of a later crystallization or devitrification of the glasses examined.

All the facts mentioned above indicate that the obsidian lava was after solidification under a state of strain, caused by the contraction resulting from the cooling of the glass. When this strain was released, minute cracks opened and only strain-free nodules were left.

Formations similar to this are described in literature as marekanites (Judd, 1886; Johannsen, 1932). Marekanites are glassy balls of obsidian, more or less rounded with concave indentations. They are left after an onion-like breaking of obsidian, the so called perlite texture. Typical of most perlite occurrences is that the parts which are split off from the marekanite balls are of the same physical state as the central marekanite, i.e. macroscopically glassy pieces of rock.

A micro-opening of glassy volcanic rocks of the kind described in the present paper is not known from old rocks of volcanic origin. It is possible, however, that such a strain-releasing breaking of volcanic glass is, and was, a comparatively common phenomenon.

Micro-cracked volcanic glasses are easily healed by silica solutions penetrating the rocks, as was found upon examination of drill-holes at Yellowstone National Park (Fenner, 1936). The foliation or onion-like parting of the obsidian rock may lead to a complete disappearance of the marekanite balls. Such a micro-cracking of the rock of obsidian lava flows, when followed by a healing of the cracks by silica, will constitute a stabilization of the strain-ridden material. The resulting rock will in most cases be subjected to a devitrification, changing the glass to a more or less holocrystalline rock. After devitrification has taken place it will be very difficult to see any difference between an originally holocrystalline lava and the devitrified obsidian. The devitrified rock will still be characterized by lamellar flow structure, representing the original distribution of microlites and crystallites in the glass.

The micro-cracked obsidian is very sensitive to weathering, as its active surface is multiplied by the cracking as compared with the ordinary obsidian. This means that it decomposes more rapidly under favourable weathering conditions than holocrystalline lava. Such a weathering is taking place at El Fiscal within the uppermost metres of the obsidian flow, where marekanite balls are found in an argillitic matrix. The intense weathering within this region will within a few years destroy the beautiful outcrops uncovered by the building of the

Atlantic Highway, leaving a muddy mixture mingled with marekanite balls.

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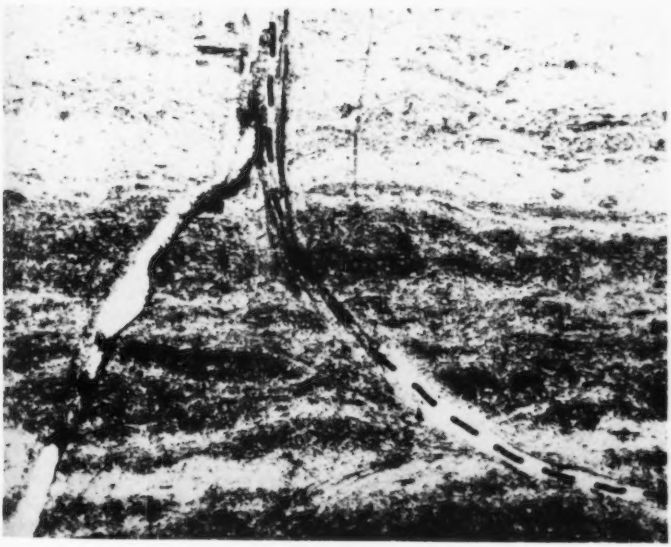
EXPLANATION OF PLATE I

FIG. 1.—Grey volcanic rock with layers of obsidian in the form of marekanite balls: El Fiscal, Guatemala.

FIG. 2.—Microphotograph of a marekanite ball (right) in contact with grey volcanic rock, showing the distribution of microlites in the volcanic glass. Ord. Light, × 16. Slide thickness, 0·1 mm.

FIG. 1



MAREKANITE FROM GUATEMALA.

FIG. 2

Facing p. 52.

Early Transverse Dykes in the Main Donegal Granite

By W. S. Pitcher and H. H. Read

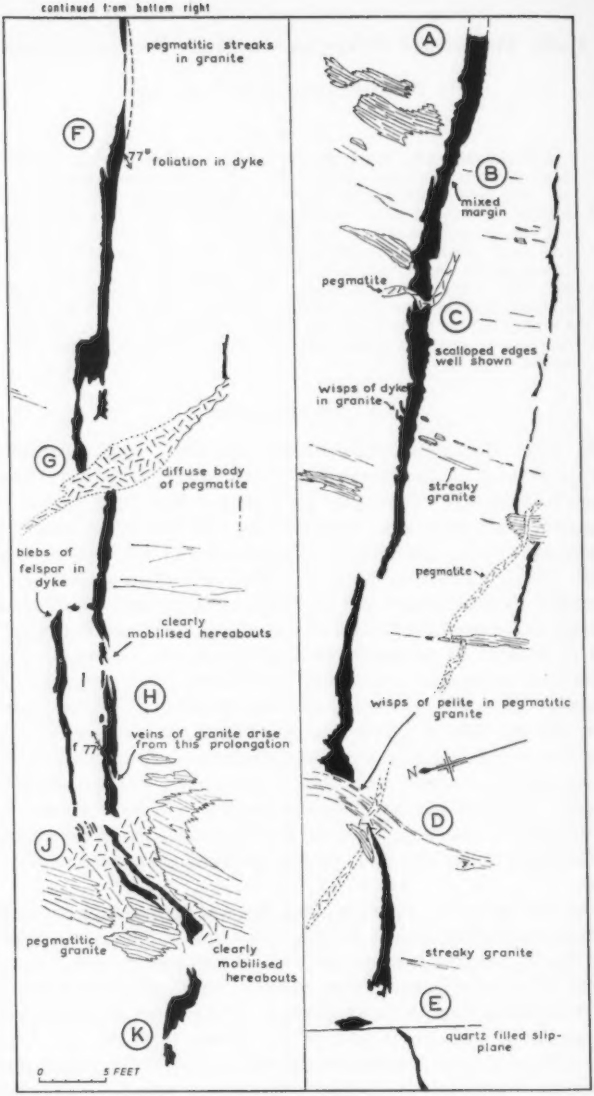
Abstract

Certain microtonalite dykes cut across the flow banding of the Main Donegal Granite in the direction of an important cross-joint, yet they and their granite host are shown to have crystallized together and in the same manner as both the darker flow bands in the granite and the pelitic xenoliths within it. It appears that an embryonic joint system existed in the Donegal Granite before consolidation and crystallization were complete. It is suggested that the early dykes represent metasomatized pelitic enclave material and are not differentation products of granitic magma. Comparison is made with the basic analogue provided by the beerbachite enclaves and dykes of the hypersthene-gabbro of Ardnamurchan.

I. Introduction

The Main Donegal Granite (Pitcher and Read, 1959) is traversed by prominent steep cross-joints which have a general south-easterly course at right-angles to both the flow banding and the marginal shear structures. Within the central area of the granite the directions of these joints show concentrations at 123° and especially 143°. Such joints may be coated with hydrothermal minerals or occupied by quartz-veins and pegmatite-aplite dykes. The latter are straight and sharp-edged and, being later than the intrusion-structures, are considered to represent the final episode in the emplacement history of the granite. At one locality, however, a couple of rather irregular dykes, with characters showing them to be of earlier formation, also cut right across the flow structure of the granite at 120° and thus near one of the standard joint-directions. A description of these dykes is given here and, from this, it seems likely that this particular fracture direction was established in embryo before the magma had finally come to rest, and before the final crystallization of the material of the dykes, the banded Main Granite itself and the xenoliths enclosed in it.

The locality of the dykes is along the Sruhanavarnis, a mountain stream running from Lough Avarnis, south-east of the Poisoned Glen, and flowing south-eastwards into the Owengweebarra above Lough Barra. The stream cascades over extensive bare rock surfaces so that the field-relations are perfectly exposed. On these surfaces banding in the granite is well shown (Pitcher and Read, 1959, pp. 271–5 and Fig. A, Plate X) and innumerable enclaves of several lithological types in orderly trains are spectacularly developed. Because this locality is so well worth seeing more explicit directions are given. The Struhanavarnis is shown, but not named, on Plate XIII of Pitcher and



TEXT-FIG. 1.—Map of early dyke in the course of the Struhanavarnis.

Read (1959), and is located in the centre of the Main Granite, between the Poisoned Glen and the conspicuous mountain of Moylenanav. It is best approached from near a bridge, 1½ miles south-west of the bend in the Doochary–Glendowan road—on this course, fording the main river is avoided. The dykes are exposed just above the lip of the hanging valley and on the north-east side of the stream course.

II. Field Observations

The granite that forms the country-rock of the dykes is typical of the north-western part of the Main Granite (Pitcher and Read, 1959, pp. 270–1)—a very coarse biotite-granite made of quartz, biotite, potash feldspar and a sodic plagioclase. It shows a poor mineral orientation trending north-east and streaks of more pegmatitic granite and a weak banding have this same direction. Ragged pelitic enclaves and wisps are abundant and are, as usual, lined up in the banding direction. Three early dykes are visible and the main part of the exposed courses of two of them is depicted in Text-fig. 1. The thin dykes running parallel to the more continuous thicker one are not further considered here; their presence suggests that more of these dykes may be found and the parallelism of the three representatives has structural implications that are dealt with later. The dykes are made of a rock darker and finer-grained than the granite host, and the main dyke, which is our concern here, can be followed continuously for over 150 feet, its width varying from a few inches to a few feet. Its detail is now described, a beginning being made at the south-eastern end (A in Text-fig. 1) where it will be encountered by anyone approaching from the Owengweebarra valley.

The first 50 feet or so of the dyke runs on a bearing at 120°. Its thickness is fairly constant, though its margins rarely match in detail of outline, but have irregular scalloped edges outside which wisps of darker dyke material lie within the light-coloured granite. Though never knife-sharp the contacts are usually fairly well defined, but on the south-western margin (B in Text-fig. 1), some 12 feet from the start of the exposure, the dyke shows a marginal variation 3 inches wide and 3 feet long, characterized by a thorough mixing of granite and dyke material. Large white feldspars lie in the darker matrix of the dyke and the contact relationship is exceedingly similar to that between granite and pelitic hornfels—its significance is discussed in a later page. Attention must be directed to the tortuous pegmatite that cuts across the dyke at locality C; pegmatite material was available, or the granite could be pegmatized, after the formation of the dyke. At another point the dyke is apparently broken apart. This first portion of the dyke appears to end abruptly (at this level of exposure) against a complex of pegmatitic granite and pelitic enclaves

(locality D); it is a fact of observation that the dyke nowhere in its length cuts through the pelitic enclaves that abound in the granite.

Beyond this interruption, the dyke starts again as a thin curving thread becoming thicker, discontinuous and irregular at locality E where, too, it is shifted on a quartz-filled movement plane. After this, the dyke changes its direction somewhat from 120°, yet further along tends to return to this standard direction as shown in Text-fig. 1. It will be later concluded that the dyke was intruded before the granitic host was finally consolidated and it may well be that the dyke was dragged out of its course by the latest stages of differential flow in the granite.

At locality F the dyke shows a steep foliation running at an acute angle across it at 77°, a structure which can be detected elsewhere along its length. Further along, in the area labelled G in Text-fig. 1, a belt of pegmatitic granite with diffuse edges obliquely crosses the course of the dyke—again it would appear that pegmatite formation outlasted the emplacement of the dyke.

Beyond this belt, the dyke shows many interesting phenomena especially at locality H. Two parallel portions, irregularly pinching and swelling and even deteriorating into thin veins, are connected across by rather discontinuous lenses. In places there are large feldspars and blebs of granitic material within the margins of the dyke which could equally well be interpreted as either xenocrystic fragments or as porphyroblasts of metasomatic origin. Further, there is one example of vein-like granite streaks arising from an almost totally enclosed lobe of the coarse granite; thus it is often difficult to establish the time relationship between the two rock types in any small part of the outcrop.

The next exposures of the dyke constitute locality J in Text-fig. 1. Here pelitic enclaves are large and abundant in pegmatitic granite; the dyke material occurs in such a discontinuous and sinuous form in the granite between two of the enclaves as to suggest that both the dyke and its host were fluxioned at the same time. The exposure of the dyke seems to end in curving irregular detached bodies, in Text-fig. 1, but actually it can be seen to continue under the waters of the Struhanavarnis.

In our opinion, the cross-cutting nature of this dark rock proves that we are dealing with a dyke; whether it be of intrusive (as we think) or of replacement origin is a matter for a discussion which does not materially affect our final thesis of pre-consolidation emplacement. But it will be shown below that the material making the dyke is exceedingly similar to the feldspathized pelitic xenoliths occurring abundantly in the deeper central portions of the Main Granite. Yet it certainly cannot be a xenolith since it runs at right angles to the flow

pattern of the Granite in which are aligned thousands of undoubted pelitic xenoliths. The course of the main dyke, in part that of a prominent cross-joint, and the existence of parallel bifurcations and of other examples with a similar course, agree with the dyke interpretation.

III. Petrography

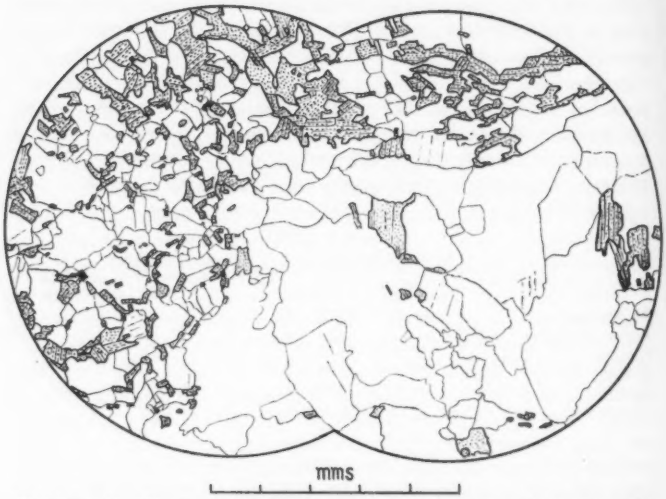
The rock of the dyke is mid-grey in colour, fine in grain, and a weak mineral orientation is often visible. In thin slice it is seen to be a microtonalite composed of plagioclase, quartz, and biotite, together with abundant dusky apatite and scanty opaques. Two modes are provided and compared with a general average of the granite host. The plagioclase is quite strongly zoned with the composition varying progressively inwards from a calcic oligoclase to a medium andesine. There are no sharp zonal boundaries but the cores are often filled with

	Microtonalite dyke 1.	Microtonalite dyke 2.	Main Granite host. (A general mode of the coarse variety of N.W. areas)
Plagioclase	49·0	51·4	27
Potash feldspar	—	—	34
Quartz	22·2	23·3	32
Biotite	27·8	24·9	6
Apatite	0·7	0·4	} 1
Opaques	0·2	tr.	

epidote-orthite grains. The brown biotite is abundant in small plates, often enclosed in the feldspars. The foliation recorded in the field is shown in thin slice by the alignment of these biotite plates which are also plastered on some parts of the contact and follow the small irregularities of it. The general texture of the dyke rock is granular with a marked similarity to that of a coarse hornfels and, therefore, to that shown by many of the feldspathized pelitic rafts in the granite. In some slices of the margin there occur scattered large homogeneous oligoclase feldspars and aggregates of quartz, plagioclase, and biotite that look like fragments of the granitic country rock though they are now completely intergrown with the groundmass. These are the questionable xenocrysts or porphyroblasts and the closer examination still fails to decide their origin.

The granite in contact with the dyke is a very coarse biotite-granite, rather abnormally rich in apatite and containing an unzoned oligoclase. The contact is quite well defined but usually rapidly transitional, rather irregular, and lobed. There is no suggestion of a chilled edge, and the contact is marked simply by a falling off in the amount of biotite, a certain contrast in plagioclase composition, the absence of

potash feldspar on the dyke side, and the great change in grain size (Text-fig. 2). Indeed, in our opinion, it is the same biotite of the same age that occurs on both sides of the contact, and the plagioclase is clearly changing to a composition nearer to that of the granite. Thus by examination of thin sections alone, it would be scarcely possible to determine which rock was intrusive, though there is the fact that the biotite of the dyke in places follows the irregularities of the contact.



TEXT-FIG. 2.—Contact of early dyke with granite host. Biotite stippled, quartz and feldspars simply outlined. Shows the continuity of the crystal fabric across the contact and, at the top of the figure, the coarsening of the biotite in a wisp-like protuberance into the host.

Indeed, hand specimens and slices showing this contact would reasonably be taken, away from the field, as being those of a fine grained feldspathized xenolith in the coarse granite. Especially in thin slice the contact is exactly similar to those between this granite and its pelitic enclaves. Again there is one minor difference, only the plagioclase of the dyke is distinctly zoned. The granite immediately adjacent to the dyke is in no way different from the coarser rock type of the Main Granite as a whole, so that the minor intrusive cannot itself be the cause of the textural merging. More is said on these topics below.

IV. Discussion and Implications

It seems reasonable to suggest that the dyke material came into place before the granite was entirely solid. The mixing at contacts,

the merging of the textures, the intrusion of granitic material into the dyke and of dyke material into the granite, the detail of the edges and the sinuosities of the dyke, the shift of the dyke out of a single alignment, are all points in favour of this interpretation.

The direction followed by important portions of this early dyke is in agreement with one set of cross-joints and it is a reasonable deduction that it followed the early representatives of such joints. It appears then that an embryonic joint-system existed in the Donegal Granite before consolidation was complete. Grout and Balk came to similar conclusions regarding joints in both the Boulder Batholith and the Snowbank Stock of Montana (1934 B, p. 889 and 1934 A, p. 629). Referring to a fragmented lamprophyre they state that " in some places even the syenite from the walls of the dykes injects the dykes showing that the fractures and lamprophyres developed early, long before the final consolidation of the mass ".

The condition of the granite when the dyke material came in is a matter of much interest. We have already seen that at least the final episodes in the crystallization of both dyke and granite have occurred simultaneously. It also seems that the mineral assemblage of the dyke was adjusting itself to that of the granite host in much the same way as a xenolith immersed in granite magma, the minerals of both tending to reach equilibrium in what was effectively the same physico-chemical system.

This similarity of cooling history and crystallization can be extended to two other pairs of rocks in the Main Granite. First, the feldspathized pelitic rafts enclosed in the Granite show a similar crystallization to that of the dyke-material. This crystallization was clearly coeval with that of the enclosing granite. Secondly, dark bands show the same crystallization as the light bands in the banded granite and there is no reason why the bands should not have crystallized together. From these observations, it can be proposed that the crystallization now shown by the banded granite, the xenoliths in it and the dykes cutting it is *all one and the same crystallization.*

If this proposal is accepted then it is necessary to inquire whether this crystallization is the only one that these rocks have undergone or whether it is late and superimposed on an earlier crystallization that may have occurred at different times in the different rocks. Whatever changes did occur were clearly of a widespread and wholesale character in no way localized to the vicinity of the dyke. Despite the interlocking of the textures, the wrapping of the marginal biotite flakes around some large plagioclase crystals abutting into the dyke seems to show that the granite was largely crystalline on intrusion of the dyke. Thus, two phases of crystallization are envisaged in the host, a second merely modifying the first, a view agreeable to recent statements on the development of granite textures by Tuttle and Bowen (1958). As there

is no evidence that the second phase was a separate and wholly later event it seems reasonable to consider that the whole crystallization of the dyke was coeval with the later part of the continuous cooling history of the host.

The early dykes have a similar composition to the finer grained and more cleanly intrusive felsite dykes which cut the metasediments of the *roof zone* of the Main Granite (Iyengar, Pitcher, and Read, 1954) with the same north-westerly trend. These dykes may well represent the Sruhanavarnis intrusions at higher levels. We may suggest that at lower levels the dyke-material may have been produced by the mobilization of metasomatized pelitic xenoliths deep in the granite.

This suggestion that the early dykes may be mobilized enclaves recalls the views of M. K. Wells (1951, 1954) on what appears to be an exact basic analogy in the hypersthene-gabbro of Ardnamurchan. This gabbro contains innumerable granoblastic basic masses that are certainly xenoliths; the gabbro is just as certainly cut by a few dykes of beerbachite that is identical with the rock of the basic enclaves. A figure of these dykes presented by Wells (1954, fig. 9, p. 391) is a replica of the geometry of the Sruhanavarnis early dykes. Both the Ardnamurchan and Donegal examples show the same kind of marginal mixing. Wells concludes that the gabbro was still hot when the dyke-material was crystallizing and that " in the absence of convincing field-data it is difficult to distinguish between an injected beerbachite and a xenolithic basic granular hornfels ". It will be recalled that the same difficulty arose in the Odenwald, the original locality of the beerbachites. Chelius followed by Rosenbusch described them as dykes cutting through the Frankenstein gabbro, Klemm has shown that they are not dykes but basic hornfels blocks enclosed in the gabbro (for a discussion and references, see MacGregor, 1931). As Wells has remarked, these diverse views may yet be reconciled. As the Sruhanavarnis dyke-material may be metasomatic xenolithic in origin, then early dykes may not be in any sense differentiation products of a granitic magma. Wells preferred also to regard the beerbachite magma as derived " from partially fused and mobilised xenolithic material . . . squeezed into the dykes in a crystallizing state, finally consolidating under the temperature and pressure conditions appropriate to the ' pyroxene-hornfels ' metamorphic facies " (*loc. cit.* p. 391). The possible equivalence of some dykes and some xenoliths should be borne in mind in the erection of elaborate genetic schemes for the igneous rocks of any area.

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An Occurrence of Mugearite in Antrim

By G. P. L. Walker

(*with chemical analysis by I. S. E. Carmichael*)

Abstract

Mugearite is described for the first time from the Tertiary lava series of Antrim, Ireland.

Mugearite is a constant, if not very abundant, member of the Tertiary basalt lava series of the Hebridean region. The rock-type was first named by Harker (1904) in Skye and was later recognized by him in the Small Isles (Harker, 1908); it also composes the well-known " pale group " of Ben More, Mull (Bailey *et al.*, 1916). In the circumstances it is most surprising that mugearite has not hitherto been recorded from the Tertiary basalt pile in Antrim. It is the purpose of the present paper to place on record an occurrence of typical mugearite from south Antrim.

The rock described in this paper is well exposed in a small and long-disused quarry ¼ mile S.W. of Y Bridge, 1¼ miles N.E. of the village of Stonyford, at lat. 54° 35′ N., long. 6° 08′ W. Warren (1871) recognized that the rock was not a normal basalt and described it thus: " this rock has a smooth conchoidal fracture, and gives out a ringing sound when struck with a hammer, and resembles the rock called ' clinstone '."

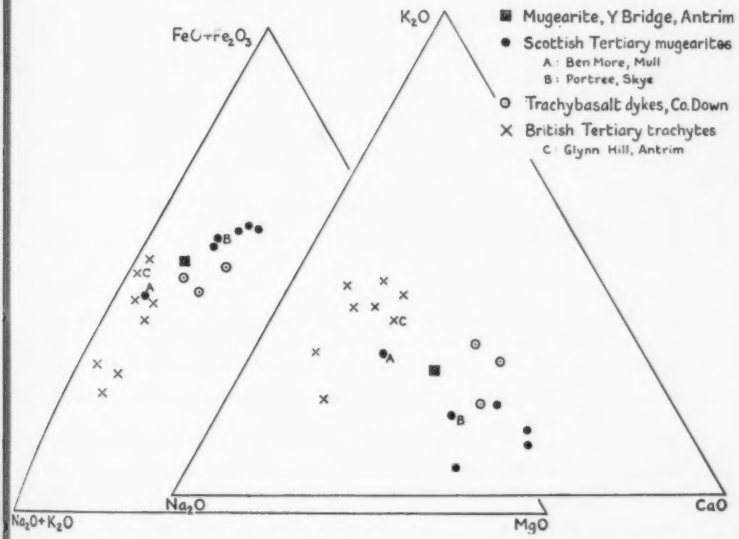
The rock is dark and basaltic in appearance, but very fine-grained, and it possesses a distinct fissility parallel to the flow structure. There is no reason for doubting that the rock in the quarry is a lava flow, and at the top of the quarry face in one place a red alteration is seen of the type found at the top of most Antrim lavas. A vesicular olivine-basalt flow which probably overlies the mugearite is seen in a road cutting and in a small opening N.W. of the quarry. Exposures in this part of Antrim are very scanty, and no attempt has been made to trace the mugearite laterally.

The position of the mugearite in the Tertiary lava succession is uncertain on account of the scarcity of exposures. The base of the lavas is exposed 3½ miles away up-dip, and the regional dip is about 5–10°; in the absence of intervening faults the mugearite must be well over 1,000 feet up in the lavas.

In thin section the rock is seen to be non-porphyritic and feldspars are the predominant constituents. Plagioclase, oligoclase in composition, forms tiny laths 0·1–0·2 mm. long with a pronounced fluxional orientation, with interstitial alkali-feldspar. Of the ferromagnesian minerals, magnetite, often in well-shaped octahedral grains, and ophitic clinopyroxene, probably augite, are equally abundant. No

fresh olivine is visible in the sections examined, but numerous green pseudomorphs probably after olivine are seen. Tiny needles of apatite also occur.

Petrographically the rock is a typical mugearite, following the definition given by Macgregor (1928). The rock differs from the type mugearite in Skye (Harker, 1904) in the absence of biotite and hornblende and in the greater relative abundance of augite. Chemically



TEXT-FIG. 1.—Chemical composition of the Y Bridge mugearite in relation to that of British Tertiary mugearites and trachytes.

(Table 1) it agrees closely with the type mugearite, although more closely with the Tertiary mugearite from Ben More, Mull, the Mull and Antrim mugearites both being atypical in possessing more than 55 per cent of SiO_2. The chemical relationship between the Antrim rock and other British Tertiary mugearites is shown on the accompanying triangular diagram, Text-fig. 1. Certain trachybasalt dykes of Tertiary age on the County Down coast (Tomkeieff and Marshall, 1935) some 25 miles S.E. of the present outcrop are also included on this diagram, together with British Tertiary trachytes. With the mugearite of Ben More, Mull, the Antrim mugearite lies near the trachyte end of the mugearite field.

TABLE 1.

	X.1.	*A.*	*B.*	*Norm of X.1.*			
SiO_2	55·9	55·76	49·24	Qz			4·9
TiO_2	1·3	1·78	1·84	Or			18·9
Al_2O_3	14·9	16·55	15·84	Ab			42·5
Cr_2O_3	nil	—	tr.	An			8·6
Fe_2O_3	5·8	3·10	6·09	Di	Wo 4·1	7·9	
FeO	4·8	6·02	7·18		En 2·5		
MnO	0·20	0·22	0·29		Fs 1·3		
MgO	1·8	1·08	3·02	Hy	En 2·0	3·0	
CaO	4·4	3·23	5·26		Fs 1·0		
Na_2O	5·0	6·28	5·21	Mt			8·4
K_2O	3·2	3·87	2·10	Il			2·4
P_2O_5	0·61	0·40	1·47	Ap			1·3
H_2O+	1·2	0·95	1·61	H_2O			2·5
H_2O-	1·3	0·80	1·08				
rest	—	0·10	0·23				100·4
	100·4	100·14	100·46				
Sp. Gr.	2·66	2·67	2·79				

X.1. Mugearite, Y-bridge, Antrim. Analyst: I. S. E. Carmichael.
A. Mugearite lava, Ben More, Mull (Bailey, *et al.*, 1924).
B. Type mugearite, Portree, Skye (Harker, 1904).

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Shear Failure of Anistropic Rocks

By J. C. Jaeger

Abstract

The two-dimensional theory of two simple generalizations of the Coulomb-Navier criterion for shear failure is developed. The first of these refers to a material with a single plane of weakness which has a different shear strength and coefficient of internal friction from the remainder of the material. In this case it is shown that failure may take place, according to circumstances, either in the plane of weakness or in planes cutting across it. The second criterion refers to a layered material whose shear strength varies continuously from a maximum in one direction to a minimum in the perpendicular direction. In this case it appears that, instead of the two directions of failure possible for an isotropic material, there is only one possible plane of failure which lies between the plane of minimum shear strength and the nearest to it of the two Coulomb-Navier planes. Numerical results are given for the case of uniaxial compression and experimental results are shown to be in reasonable agreement with them.

I. Introduction

Bott (1959) has discussed the mechanics of rocks which possess an anisotropy of strength indicated, for example, by preferred planes of weakness due to earlier faults or to cleavage or bedding. He considers the effects of low strength in such planes, but does not discuss the effects of friction across them. This problem, including friction, is a familiar one in rock mechanics [cf. Talobre (1957)] but it is equally important on a larger scale and it seems worth while pointing out that there are two simple generalizations of the Coulomb-Navier theory which are probably applicable to many tectonic situations of this type. The two-dimensional theory of these is set out below; there is no difficulty in extending it to three dimensions but the results become rather complicated.

The first case, which will be discussed in §II, is that of an isotropic material whose behaviour can be described in the ordinary way by a shear strength (or cohesion) S and a coefficient of internal friction μ, but which is supposed to have a plane (or parallel planes) of weakness with different values S' and μ' of the shear strength and coefficient of internal friction. It may be shown that for some stress-systems slip will take place in the plane of weakness, but that for others it will take place in planes determined by the Coulomb-Navier theory and these will intersect the planes of weakness. This theory has been discussed by Jaeger (1959) and shown to be in reasonable agreement with laboratory experiments on rocks containing a single plane of weakness. In the case of open joints, μ' will be a coefficient of friction in the ordinary sense, but the same theory applies. Measured values of μ' were in the range 0·5 to 0·8.

While the theory described above may be expected to give a reasonable idealization of a material with very weak joints in a particular direction, it provides a rather drastic description of an ordinary bedded rock in which a continuous variation of mechanical properties with direction might be expected. A second generalization of the Coulomb-Navier theory may be obtained with very little algebraical complication by assuming that the shear strength S of the material in a plane inclined at ω to the plane of minimum shear strength has the value

$$S = S_0 - S_1 \cos 2\omega, \quad . \quad . \quad . \quad . \quad (1)$$

so that the shear strength varies between $S_0 - S_1$ and $S_0 + S_1$. To obtain the simplest possible treatment, μ is taken to be constant; this is reasonable, since it is known that it does not vary greatly between rocks of different types, for example, Price (1959) gives a range of 0·84 to 1·38. The assumption (1) leads to some very interesting conclusions: in particular that in place of the *two* planes of failure possible on the Coulomb-Navier theory there is only *one*, which lies between the plane of minimum shear strength and the nearest to it of the two Coulomb-Navier directions.

II. Systems with a Single Plane of Weakness

Consider a two-dimensional stress-system with principal stresses C_1 and C_2, compression being reckoned positive, and $C_1 > C_2$. Suppose that the material itself has a shear strength S and a coefficient of internal friction $\mu = \tan \phi$, except across a plane inclined at β to C_1 across which the shear strength is S' and the coefficient of internal friction is $\mu' = \tan \phi'$.

The criterion for failure in any plane is that if N is the normal stress across it and τ is the magnitude of the tangential stress across it, failure takes place when τ reaches the value

$$\tau = S + \mu N, \quad . \quad . \quad . \quad . \quad (2)$$

where S and μ are the appropriate values of the shear strength and the coefficient of internal friction.

Now for a plane inclined at α to the direction of C_1,

$$N = C_m - \tau_m \cos 2\alpha, \quad . \quad . \quad . \quad (3)$$

$$\tau = \tau_m |\sin 2\alpha|, \quad . \quad . \quad . \quad . \quad (4)$$

where $\tau_m = \frac{1}{2}(C_1 - C_2)$ . . . . . (5)

is the maximum shear stress, and

$$C_m = \tfrac{1}{2}(C_1 + C_2) \quad . \quad . \quad . \quad . \quad (6)$$

is the mean of the principal stresses [see, e.g. Anderson (1951), Jaeger (1956)].

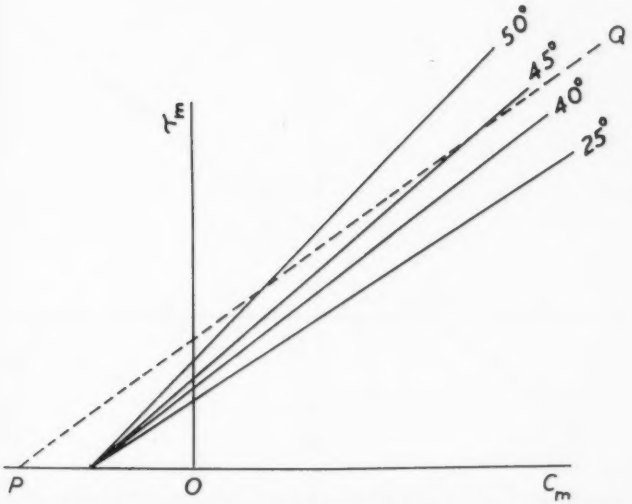
For failure in a uniform medium with shear strength S and $\mu = \tan\phi$. it is found from (3) and (4) that $\tau - \mu N$, regarded as a function of α, has its maximum value when

$$\alpha = (\pi/4) - \tfrac{1}{2}\phi, \quad . \quad . \quad . \quad . \quad . \quad (7)$$

and that, for this value of α, (2) gives

$$\tau_m = C_m \sin\phi + S\cos\phi. \quad . \quad . \quad . \quad . \quad (8)$$

Thus the relation between τ_m and C_m at which fracture takes place



TEXT-FIG. 1.—Variation of τ_m with C_m for fracture. The dotted line PQ is for failure through the material of shear strength S and angle of internal friction 45°. The full lines are for failure in a plane of weakness with shear strength $\frac{1}{2}S$ and angle of internal friction 40°: the numbers on the curves give the inclination of this plane to the direction of maximum compressive stress.

is a straight line in the C_m, τ_m plane which has an intercept $-S\cot\phi$ on the C_m-axis and inclination θ to this axis given by

$$\theta = \tan^{-1}(\sin\phi). \quad . \quad . \quad . \quad . \quad . \quad (9)$$

This is the conventional Coulomb-Navier theory.

On the other hand, for failure in a definite plane inclined at β to C_1 in which the shear strength is S' and the coefficient of internal friction is $\mu' = \tan\phi'$, substituting (3) and (4) with $\alpha = \beta$ in (2) gives

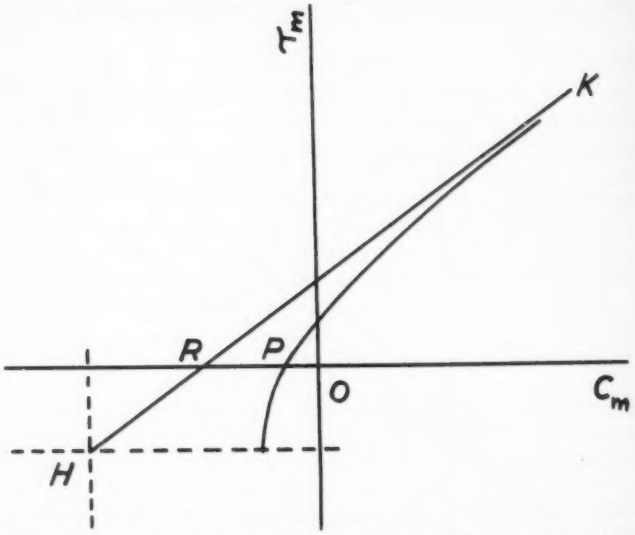
$$\tau_m \sin(2\beta + \phi') = C_m \sin\phi' + S'\cos\phi'. \quad . \quad . \quad . \quad (10)$$

This is a straight line whose intercept on the C_m-axis is $-S' \cot \phi'$ and which is inclined at θ' to this axis, where

$$\tan \theta' = \sin \phi' \operatorname{cosec} (2\beta + \phi'). \quad . \quad . \quad . \quad . \quad . \quad . \quad (11)$$

For any values of β and C_m, fracture will take place at the lower of the two values of τ_m given by (8) and (10).

Since it may be assumed that $S' < S$ and, probably, $\mu' < \mu$, the usual

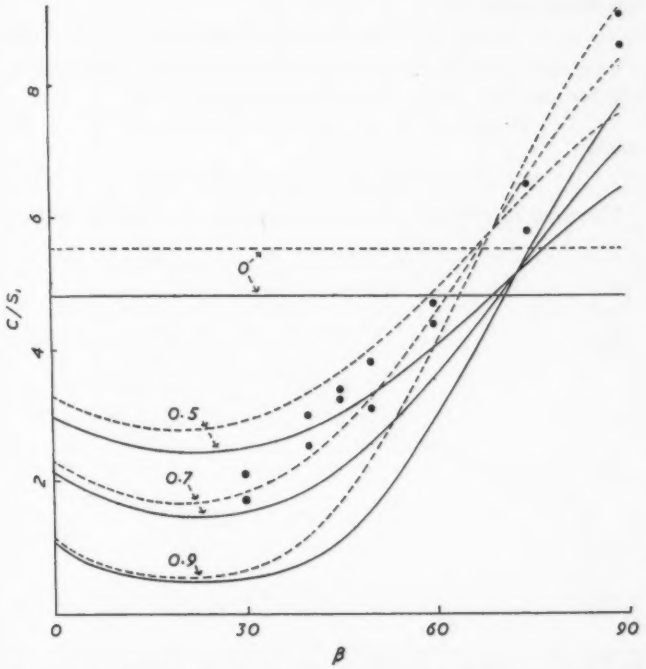


TEXT-FIG. 2.—C_m, τ_m curve for fracture of a material with variable shear strength.

situation will be that shown by Fig. 1 which is drawn for $S' = \frac{1}{2}S$, $\phi = 45°$, $\phi' = 40°$. The full lines correspond to failure in the plane of weakness for various values of β. The dotted line PQ corresponds to failure through the material according to (8). It is seen to intersect some but not all of the full lines corresponding to failure in the plane of weakness. Thus, for example, for the case in which both principal stresses are compressive, failure along the plane of weakness is impossible if, approximately, $\beta > 90 - \phi' = 50°$, and failure takes place in the material in a plane intersecting the plane of weakness: if $50° > \beta > 37{\cdot}5°$ or $0 < \beta < 12{\cdot}5°$, failure takes place in the plane of weakness for small values of C_m but through the material for larger ones: if $12{\cdot}5° < \beta < 37{\cdot}5°$, failure takes place in the plane of weakness for all values of C_m.

III. A System with Continuously Variable Shear Strength

With the co-ordinate system as before, suppose that the shear strength of the material is least in a direction inclined at β to C_1, and that it



Text-fig. 3.—Compressive strength C as a function of β for a material with variable shear strength. The numbers on the curves are values of $k = S_2/S_1$. Full curves are for $\mu = 1$ and dotted curves for $\mu = 1{\cdot}2$. Dots are experimental values for slate on an arbitrary scale.

varies according to the law (1) so that the value in the direction α is

$$S = S_1 - S_2 \cos 2(\alpha - \beta), \quad . \quad . \quad . \quad . \quad (12)$$

where S_1 and S_2 are constants. For the present it will be assumed that $0 < \beta < \frac{1}{2}\pi$ and $0 < \alpha < \frac{1}{2}\pi$. Using (3), (4) and (12) in (2), the criterion for failure becomes

$$(\tau_m + S_2 \sin 2\beta) \sin 2\alpha + (\mu\tau_m + S_2 \cos 2\beta) \cos 2\alpha = S_1 + \mu C_m. \quad . \quad (13)$$

For any given value of τ_m, the left hand side of (13) has its maximum value when

$$\tan 2\alpha = \frac{\tau_m + S_2 \sin 2\beta}{\mu\tau_m + S_2 \cos 2\beta} \quad . \quad . \quad . \quad (14)$$

and this maximum is

$$[(\tau_m + S_2 \sin 2\beta)^2 + (\mu\tau_m + S_2 \cos 2\beta)^2]^{\frac{1}{2}}. \quad . \quad . \quad (15)$$

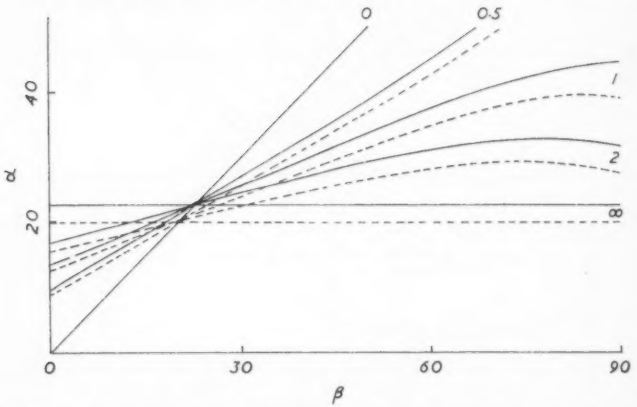
Thus by (13), failure will take place for this value of τ_m if the quantity (15) is equal to $S_1 + \mu C_m$. This condition may be written

$$\mu^2(C_m + a)^2 - (1+\mu^2)(\tau_m + b)^2 = S_2^2 - b^2(1+\mu^2) \quad . \quad (16)$$

where

$$a = S_1/\mu,\ b = S_2\ (\sin 2\beta + \mu \cos 2\beta)/(1+\mu^2) \quad . \quad . \quad (17)$$

This is an hyperbola in the C_m, τ_m plane, Fig. 2, whose centre H is



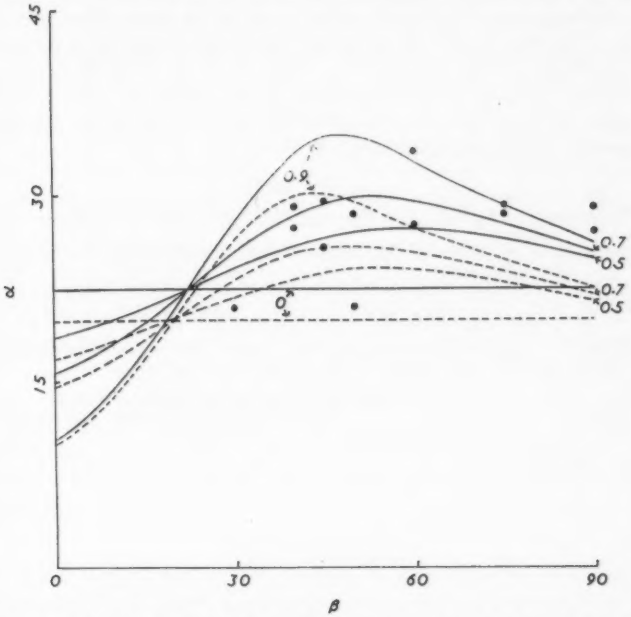
TEXT-FIG. 4.—Variation of α given by (14) with τ_m/S_2 and β. The numbers on the curves are values of τ_m/S_2. The full curves are for $\mu = 1$ and the dotted curves for $\mu = 1{\cdot}2$.

at $(-a, -b)$ and whose asymptote HK is inclined to the C_m-axis at $\tan^{-1}(\sin\phi)$. This asymptote cuts the C_m-axis in the point R, $C_m = -[S_1 - S_2 \sin(2\beta+\phi)] \cot\phi$. The point P at which the curve (16) cuts the C_m-axis is $C_m = -(S_1 - S_2)\cot\phi$, and is independent of β.

The significance of these results may most easily be seen by considering simple stress-systems. For example, Fig. 3 refers to uniaxial compression $C_m = \tau_m = \frac{1}{2}C$, $C_2 = 0$. In this figure the value of C/S_1 necessary for fracture is plotted against β for various values of $k = S_2/S_1$: the dotted curves are for $\mu = 1{\cdot}2$ and the full curves for $\mu = 1$. The dots are measured compressive strengths on an arbitrary scale for cylinders

of slate cut at various angles β to the cleavage. Clearly, they show a trend which is in qualitative agreement with the theory.

Considering, next, the direction of failure, Fig. 4 shows the variation of α with β and τ_m/S_2 given by (14), the full curves being for $\mu=1$ and the dotted curves for $\mu=1{\cdot}2$. It has to be remembered, however, that



TEXT-FIG. 5.—Variation of the angle of fracture α with β for the case of unconfined compression. The numbers on the curves are the values of S_2/S_1. Full curves are for $\mu = 1$ and dotted curves for $\mu = 1{\cdot}2$. Dots are experimental values for slate.

τ_m is not now an independent quantity but is the value for failure determined from (16) with the given values of C_m and β. For example, in the case of unconfined compression in which Fig. 3 gives the variation with β of the value of $C/S_1=2\tau_m/S_1$ necessary to cause fracture, Fig. 5 shows the variation of the angle of fracture α with β for various values of μ and S_2/S_1. In all cases the direction of fracture is seen to have moved away from the Coulomb-Navier direction towards the direction of least shear strength. This displacement rises to a maximum and then falls off as $\beta \to 90°$. It is extremely difficult to measure angles of failure in unconfined compression because the original

failures are usually masked by secondary effects, but the dots in Fig. 5 show some experimental values for slate which exhibit a trend roughly in accordance with the theory.

Finally, the question of negative values of α must be considered. Clearly, no generality is lost by the assumption made earlier that $0<\beta<90°$ and all possible cases will be included in the range $-90°<\alpha<90°$ of α. If α is negative, $\tau_m \sin 2\alpha$ in (13) is to be replaced by $\tau_m|\sin 2\alpha|$ and it is easy to see that, if $S_2>0$, the left hand side of (13) attains greater values for $\alpha>0$ than for $\alpha<0$. Thus the criterion for failure is always satisfied first for positive values of α and there is only one plane of failure instead of the two conjugate planes possible in the Coulomb-Navier case $S_2=0$.

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Deformation Features in Cataclastic Quartz

By Glyndwr Jenkins

(PLATES II, III)

Abstract

Petrofabric analysis has been carried out on Carboniferous rocks involved in thrust masses of the Benton system in Pembrokeshire. Evidence is contributed, which indicates stages of fracturing and crystallographic control in the deformation of individual quarts grains.

Introduction

The present study is a part of an investigation of the Pre-Cambrian of South-west Pembrokeshire and the adjacent Carboniferous rocks involved in the Benton Thrust System. It was suggested by Dr. N. Rast who kindly read the manuscript of this paper.

At the western extremity of this thrust system a tectonic sliver of the Millstone Grit Series is separated by a major thrust from the Pre-Cambrian to the south. To the north, the tectonic sliver again has a thrust contact with the Coal Measures. Lithologically, the sliver consists of dark blue mudstones, thin shale bands and sandstones, the mudstones and shales alternating with dark, sugar-textured grits up to 1 foot thick. A quantitative attempt was made to determine the amount and type of post-Carboniferous deformation produced by the thrust movement. A microscope study of the grit beds revealed deformation features in individual quartz grains which warrant an independent description.

Observations

The grit band shows no megascopic deformation other than the development of a-c joints with minor oblique *l k o* joints. Mineralogically, the rock consists of sutured quartz grains with a few grains of plagioclase, perthite, myrmekite, and quartzite. The quartz grains generally vary in size from ·75 mm. to 1·25 mm. except along some irregular fractures within the rock where they are brecciated into much smaller particles. Under the microscope, the quartz grains all exhibit undulose extinction and have narrow recrystallized borders. Recrystallization has also taken place along micro-fractures in the rock. Petrofabric analysis shows preferred grain orientation (Text-fig. 1). The diagram shows a discontinuous cleft girdle with a well developed maximum I. Observation of the local megascopic structures indicates that the maxima coincides with the direction of movement of slickenside markings. It is suggested that the cleft girdle was developed during folding and that the formation of maximum I is the result of a later thrusting parallel to the direction of tectonic transport of the folding.

Axial Distribution Analysis.

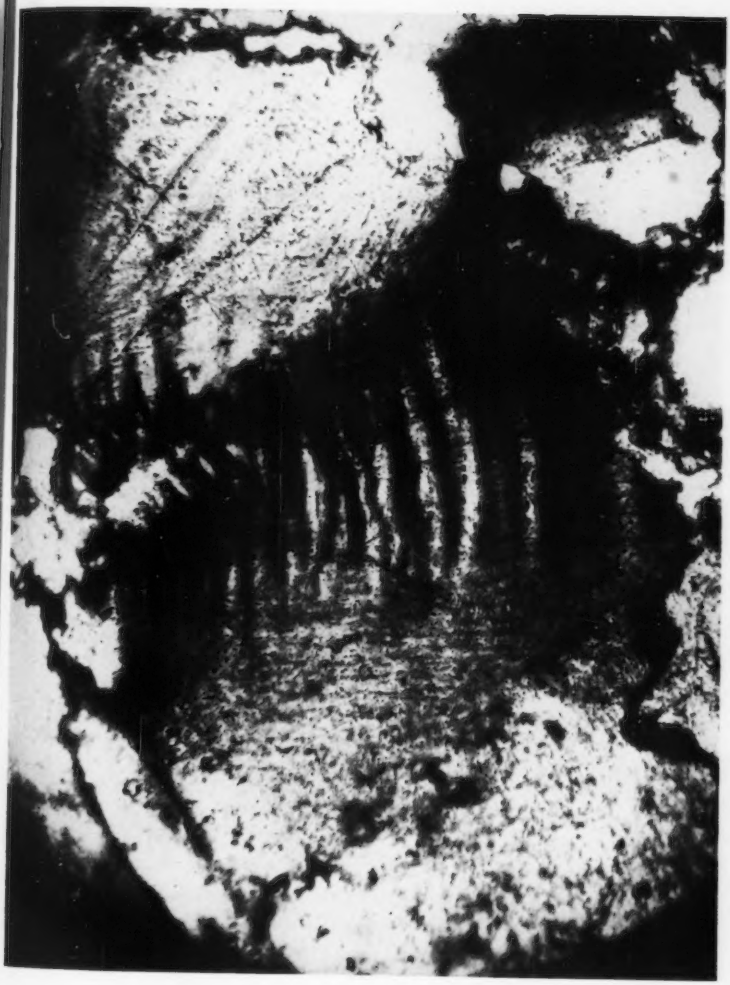
In this paper the term Bohm striations will be confined to the lines of bubble inclusions and the lens shaped areas of undulose extinction will be referred to as deformation lamellae.

Two nearly adjacent grains of different optical orientation were selected and detailed universal stage investigation was carried out with the help of methods used by Hietanen (1938). In the study of the

TEXT-FIG. 1.—210 Optic axes of quartz. Contours 6–4–2 per cent.

grains under consideration the position of the optic axes of the deformation lamellae were measured, and an optic axial map was made for each of the grains.

In Grain 1 (Plate II) two distinct sets of fractures are developed. One set consisting of two fractures, which on the photographs have a horizontal position. The lower fracture of the two is entirely recrystallized and is recognized by the slightly different extinction of quartz along it. The upper one still appears as a distinct fracture in the centre of the grain, and a series of deformation lamellae are developed sub-perpendicular to the trace of the fracture, which, when followed to the left appears as an open crack filled with crystallized quartz. The second set of fractures in Grain 1 is seen as lines of bubble inclusions or the Bohm striations of certain authors trending approxi-



CATACLASTIC QUARTZ.

Facing p. 74.

mately N.E.–S.W. in the upper part of the photograph. These are not confined to the margins of the grains as are the horizontal fractures, but have a constant orientation throughout the slide.

The N.N.E.–S.S.W. fractures of Grain 1 are parallel to the approximately N.N.W.–S.S.E. fractures of Grain 2 (Plate III), which are

TEXT-FIG. 2.—Distribution of optic axial direction in Grain 1. The "light" deformation lamellae seen in Plate II are stippled, the "dark" deformation lamellae alternating with them are unstippled. Broken lines indicate zone of unruptured strain shadows.

clearly seen to displace the deformation lamellae. In Grain 2 as in Grain 1 there is also present the association of deformation lamellae with approximately horizontal fractures which cannot be traced beyond the grain boundary.

The other deformation features present are the lines of inclusions or the true Bohm striations, which are developed only in areas where deformation lamellae are present in the grains, these striations having a different orientation when traced from one grain to the other.

Deformation Lamellae.

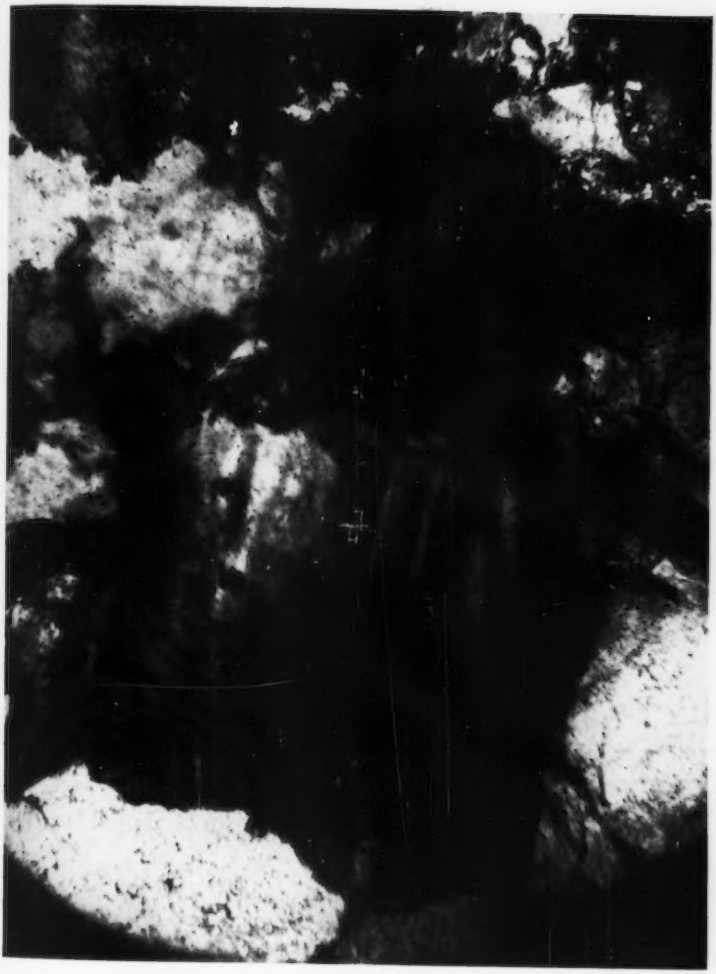
These appear to develop as a series of narrow lens-shaped lamellae which maintain a regular arrangement except where they are displaced by late micro-fractures as in Grain 2 and as to the left of the photograph

TEXT-FIG. 3.—Distribution of optic axial directions in Grain 2. Broken lines indicate the late fractures.

in Grain 1. Examination with the universal stage suggests that these bands are in the form of plates with lenticular cross-section rather than needle shaped fragments.

The association of deformation lamellae and partly recrystallized fractures nearly parallel to the basal pinacoid agrees with Hietenan's findings in the Finnish Quartzite.

Detailed investigation of the deformation lamellae both individually and collectively has revealed the following features.



CATACLASTIC QUARTZ.

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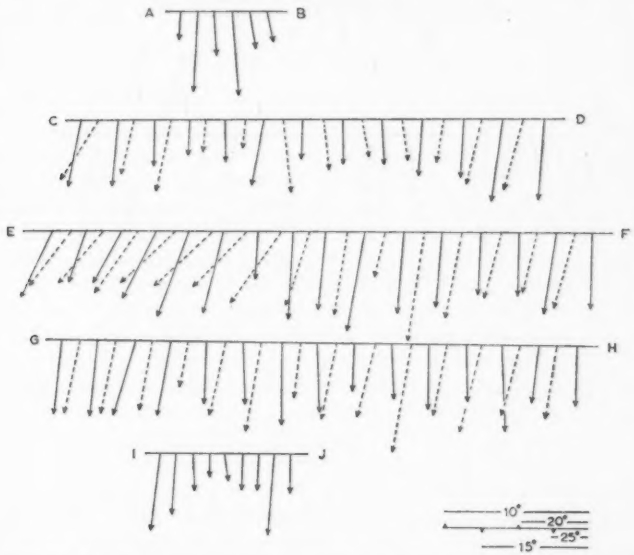
(1) The deformation bands in Grain 1 and in some parts of Grain 2 have a curved outline which is followed by a corresponding slight change in the direction of the optic axis when plotted for the whole length of individual deformation lamellae. (Text-figs. 2 and 3.)

(2) Plotting of the orientation and plunge of the optic axes in " dark " and " light " deformation lamellae and the construction of sections approximately normal to the optic axes (Text-figs. 4 and 5) show that :—



TEXT-FIG. 4.—Diagrammatic representation of the optic axial directions and their inclination in Grain 1.

Optic axes within " light " deformation lamellae.
Optic axes within " dark " deformation lamellae. The length of the arrows is proportional to their inclination.

(*a*) The variation in the direction of the optic axes of the dark deformation lamellae conforms to a " wave " pattern.

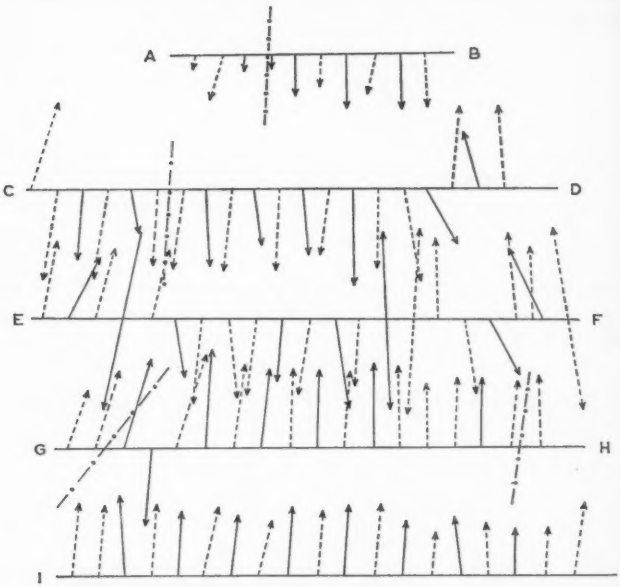
(*b*) Similar but more marked variation in the optic axes of the " light " deformation bands.

(*c*) A variation in plunge is common to both " dark " and " light " deformation lamellae.

(*d*) Both the plunge and orientation readings differ slightly in adjacent lamellae.

Early or True Bohm Striations.

These are found only within the areas of well developed deformation lamellae. The close association of these two deformation features is further illustrated where, as in Grain 1 the deformation lamellae have a curved outline, then the Bohm striations resemble the radii of a



TEXT-FIG. 5.—Diagrammatic representation of the optic axial directions and their inclination in Grain 2.

Optic axes within the " light " deformation bands.
Optic axes within the " dark " deformation bands.
Position of late fractures across the line of section.

circle with the two outside radii subtending an angle of 12° to each other, the striations maintaining a position normal to the deformation lamellae. The true Bohm striations differ from the pseudo-Bohm striations in that they are confined to within the grain boundaries and have a different orientation in other grains showing similar deformation features. The displacement of the striations by the pseudo-Bohm striations and the N.N.E.–S.S.W. fractures indicate that they belong to an early stage in the fracturing of the quartz grains.

Late or pseudo-Bohm Striations.

The following generalizations can be made about the late Bohm striations:—

(1) These displace both the early Bohm striations and the deformation lamellae.

(2) In some grains they appear as lines of liquid inclusions while in other grains they are seen as micro-fractures.

(3) They are not confined to grain boundaries and maintain a constant direction throughout the rock, subtending a varying angle with the optic axes of the grains through which they pass.

Mechanics of Deformation

Translation on (0001).

The deformation features described are similar to the features recorded by Hietanen (1938); and many of the observations support the Frontwendug hypothesis proposed by Mugge (1896) which was subsequently amplified by Hietanen's work on the Finnish Quartzites.

According to her hypothesis, strain imposed by local intergranular stress patterns has produced glide folding parallel to the basal pinacoid (Text-fig. 6*a*), resulting in the production of unruptural strain shadows as seen in the upper and lower parts of Grain 1. This mechanism imposed a control on the orientation of the optic axes, resulting in the "wave" pattern seen in the sections constructed normal to the deformation lamellae. Continued stress eventually fractured the (0001) along planes parallel to prism and rhombohedral faces with the development of well defined deformation lamellae (Text-fig. 6*b*).

In terms of Hietanen's hypothesis the variation in the plunge of the optic axes is explained by either or both of the following mechanisms.

(*a*) The local intergranular stress pattern may produce a torque effect on the basal pinacoid either during the early plastic stage of deformation or after the break up of the basal pinacoid, the latter mechanism is shown on Text-fig. 6*c*.

(*b*) Fracture along the prism and rhombohedral faces will produce planes slightly inclined to the plane containing the optic axes. Movement along these fractures would produce a variation in orientation as well as in the plunge of the optic axes of the deformation lamellae.

In her hypotheses, Hietenan suggested that the Bohm striations were traces of the original position of the basal pinacoid and were earlier than the development of the deformation lamellae or ruptural strain shadows. Investigation of the quartz grains under discussion, however, indicate that the early or true Bohm striations were later than the development of the deformation lamellae. During the discussion of these observations with Dr. N. Rast and Mr. A. L. Harris, an alternative

TEXT-FIG. 6.—Interpretive diagram illustrating the relationship between the segments of the displaced basal pinacoid within quartz as deduced from the optic axial direction. The arrows represent the optic axes within the segments.

mechanism was evolved which would also impose a " wave " control on the variation of the optic axes.

Translation on Prism and Rhombohedral Faces.

The intergranular stress pattern is envisaged as having produced stress acting parallel or sub-parallel to the optic axes of the grains. In the region of maximum stress concentration towards the central

part of the grain, fractures would be produced along prism and rhombohedral faces following plastic deformation; movement of the resulting lensoid shaped deformation lamellae would produce a variation in plunge and orientation of the optic axes, giving an appearance of folding the basal pinacoid. This mechanism of folding of the basal pinacoid is analogous to the gleibrett folds of Schmidt (1932), and the control on the " wave " variation would be imposed by a restriction of space after limited movement of the deformation lamellae along the prism and rhombohedral fractures (Text-fig. 6*c*). Continued stress parallel to the optic axis of the grain is suggested as resulting in the development of " pseudo cleavage " or true Bohm striations approximately at right angles to the deformation lamellae; their development consistantly normal to the deformation lamellae being controlled by the existence of the basal pinacoid as a possible fracture plane within the grain. In Grain 1, further stress has slightly folded the deformation lamellae and eventually fractured the grain along the approximately E.–W. fracture which displaces the deformation lamellae and the true Bohm striations. Where, as to the left of Grain 1 (Text-fig. 1) the fracturing has become intense, the deformation lamellae are further broken down into smaller fragments with a marked variation of the optic axes when compared with the parent grain.

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EXPLANATION OF PLATES

PLATE II.—Photomicrograph of Grain 1. The deformation lamellae slightly bent, and are confined to the central portion of the grain. The early or true Bohm striations are seen running horizontally approximately normal to the deformation lamellae × 400. (Facing p. 74.)

PLATE III.—Photomicrograph of Grain 2. The late fractures are well developed and displace the deformation lamellae × 400. (Facing p. 76.)

CORRESPONDENCE

ULTRABASIC PILLOW LAVAS FROM CYPRUS

SIR,—Mr. R. A. M. Wilson's remarks on the Cyprus Diabase (*Geol. Mag.* 1959, xcvi, 257, following my letter of 1958), are timely. If my original explanation has been challenged, the reasons and an alternative must needs be discussed ; and I am now able to clarify some points and to add some new information.

Mr. Wilson may not have known that I had been compiling a small-scale map showing the main axes of inflection and dips for the whole Diabase region ; but the data have been utterly lost since 1953, before completion. Subject to the lapse of time, I recall that a few—at least two—major " turn-overs " on axes could be located ; but that they were extremely sharp. The change of dips from easterly to westerly across the formation is not in doubt ; and my original suggestion was that of an anticlinorium with the great plutonic intrusions in the core.

Now the Diabase is not a piece of parish geology ; it is one of the striking features of the island. It forms saw-edged ridges and precipitous canyons 3,000 feet deep ; and within the girdling ring of lavas, extends at least 35 miles from east to west across strike, with a maximum breadth of 18–20 from north to south. The peaks of Madhari and Papoutsa to the east of Mt. Troodos are very nearly as high. Everywhere the broad regional picture is the same: sheets of very uniform types of rock a few inches to a few feet thick, picked out by their different reactions to weathering ; with variable but rather high dips, sometimes resembling a hillside of Welsh slate or Arenig lavas, and generally beautifully exposed. There is no doubt that they antedate the Pillow Lava Series and the plutonic bosses, whose eroded tops, surrounded by Diabase, are at a considerable height above the highest recorded lava.

The factual picture that Mr. Wilson and his colleagues give in paras. 3, 4, and 5, of his letter, does not differ from my original findings. On passing from the Diabase to the Basal and higher group of lavas, the feeder dykes which are injected through and are concordant with the Diabase become so numerous that it is hardly possible to tell which is which or to draw an exact boundary ; for this reason this composite material, which retains the orientation of the regional Diabase structure, is given the name of " pack " on my field maps, from analogy with a pack of cards. In vertical section there is a wide transition zone between lava proper and Diabase of a fairly consistent character ; while the later lavas may sometimes be seen lying unconformably across the upturned edges of the Diabase sheets. I agree also that the low-grade regional metamorphism and epidotization of the Diabase may well be due to action from the plutonic intrusions which have also contributed some notable dykes into the Diabase near their margin. Mr. Wilson remarks that the structural pattern of the Basal Group Dykes . . . is precisely the same as that of the Diabase sheets. Of course it is ; the dykes are lit-par-lit injections, as can be seen in some beautiful exposures, and have been controlled by the pre-existing " grid ".

There are not many possible explanations for this regional structure in the Diabase. At a distance the formation resembles sedimentary beds ; it is clearly not plutonic, nor is it a " layered intrusion " ; nor can one suggest that its sheeted units have been developed by shearing, although there is in fact plenty of shearing to be seen locally. If considered as a folded series, the Diabase with its supplementary dykes (excluding feeders to the marginal lavas), must be of the order of 10,000 feet thick, which seems reasonable enough. But if I read Mr. Wilson's paragraph (8) correctly, he believes the formation to represent the original inclinations of successive (concordant) dykes. If so, into what framework were they injected ? There is no trace of any pre-existing rock. Each dyke would have to solidify before its neighbour came in alongside of it, so that some marginal chilling would occur. Is it in any way conceivable that such a process could take place, following a definitely

controlled geometrical pattern, over a continuous width of 35 miles, each thin dyke coming in consecutively like the leaf of a book? And how was the heat dissipated so as to provide sheets instead of a continuum? Obviously the mechanism is impossible if considered on a regional scale. All that is clear is that the rock must have been emplaced in discrete sheets which were able to cool more quickly than a solid plutonic mass.

So far from there being a simple reconnaissance, several very long sections across the Diabase strike were examined in close detail. I nowhere found any intercalated sediments or a material departure from a pretty uniform character—the idea being to look for a marker horizon—and I saw no chill-margins; but these in any case might have been indecipherable, being at the parting planes of the weathered sheets. What I did find in some sheets were little round concretions of chloritic matter like peas, which may have represented amygdales. It is possible that a bore-hole core in completely unweathered rock would show evidence of marginal chilling of the sheets. It is difficult to imagine a book of thin dykes all in perfect contact with each other over a thickness of 35 miles, and no signs of matrix or margin.

In 1958 I drew attention to a possible analogy in Fuerteventura (Canary Islands), which has been insufficiently studied. Another comparison is now available with the far greater and more widely spread Ventersdorp System of basic lavas in South Africa. The later Ongeluk lavas are similar but much smaller. The Ventersdorp lavas may attain a thickness up to 15,000 feet, and are much less altered than the Cyprus Diabase; unlike the latter, they are not deeply dissected and exposed. There are very rare instances of intercalated sediment in this great formation; normally there are no markers, but some parts are highly amygdaloidal and others not so. There are, however, appreciable regional dips, while around the Vredefort Ring the whole formation has been inverted and overturned. Excellent examples of chill-margining at the interfaces of flows have been obtained in shaft-sinking; but these zones may be only an inch or less wide. At surface individual flows weather out to make little scarplets with visible partings that look rather like shears but which do reflect the dip of the rocks. The surfaces of a flow will hardly ever be identified on a weathered outcrop; but a succession of tilted flows gives a diagnostic topography. It is quite unnecessary in many lava systems that the characteristics mentioned in Mr. Wilson's paragraph 6 should invariably be present, particularly with basic material. I agree that the strain-slip type of small shear seems to be lacking in the Diabase; but the surfaces between the narrow sheets may have allowed enough slip to permit bending of such a rigid rock, like the leaves of a spring. While the dips are high, the amplitude of the "folds" is large; the term "highly folded" could apply to major structures but not to minor phenomena within those structures. It is, as Mr. Wilson observes, a very competent rock; and liable to break rather than bend.

In summary, I can find no strong reason for discarding the idea that the bulk of the Diabase, which amounts to some hundreds of cubic miles, is made up of ancient flows now showing high angles of dip, with a regional strike that is dominantly north and south. The only alternative so far suggested is that it is a pack or book of thin successive dykes or sheets which conform to a geometrical pattern resembling systematic folding. This seems impossible owing to the enormous extent of the phenomenon and the absence of any framework of rock into or through which they could have been injected. If a large proportion of the formation is an intrusive rock, into what is it intruded? The very restricted width of these sheets—over a minimum width of 35 miles—suggests a rapid succession of effusive lavas. In a tremendous system of tens of thousands of successive *dykes* some would have been much wider than a few feet and of coarser texture. The problem of heat dissipation in such a mass would have led to plutonic conditions. The comparison with the Ventersdorp lavas of South Africa certainly suggests that there is nothing unreasonable in the view that the Cyprus Diabase is a system of sharply folded ancient lavas which have been intruded and metamorphosed by plutons and their related dykes, and again pierced round the rim by later lavas and

covered by their flows. The view that the whole of the Diabase is a " book " of oriented intrusive dykes seems to me to provide insuperable difficulties, but such possibilities must be considered. However, since the formation is dissected vertically for several thousand feet, the thin Diabase sheets, if they are intrusive, might well show systematic differences in texture between points 1,000 feet and 5,000 feet above sea-level. It does seem important to record and analyse the structures so generously presented by Nature, before venturing farther. The network of motor-roads now available would make the recovery of this information a not-too-troublesome task. In the meantime I believe that my original description of " Folded Diabase " should be retained until there is stronger evidence against it.

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16th August, 1959.

CARBONIFEROUS LIMESTONE–NAMURIAN JUNCTION IN IRELAND

Sir,—The nature and extent of the non-sequence between the Carboniferous Limestone and the overlying Namurian in Ireland is rapidly being evaluated by many workers, so that a complete synthesis of the phenomenon may shortly be expected. I think, therefore, that it may be worth while to record relevant results derived from a preliminary survey of the Taur-Meelin anticline near the Co. Limerick–Co. Cork border, recently carried out.

In this region the top of the Carboniferous Limestone, seen in a series of dome-like outcrops between Taur and Meelin, is cherty to a depth of about 6 feet below the Namurian junction. Immediately above the limestone occur 60 feet of shales, corresponding to the Clare Shales of Co. Clare and Co. Limerick, above which the facies changes and sandstones and siltstones appear in the succession. West of the village of Taur, poorly preserved *Homoceras* sp. has been collected from the shales 20 feet above the top of the Carboniferous Limestone. It is clear, therefore, that lower beds in these shales are of *Homoceras* age so that the *Eumorphoceras* zones are probably absent here.

Above the Carboniferous Limestone there occur beds of thickness of the order of 1,750 feet, which include coal seam(s?) near the top. Spoil heaps from the old collieries yield shales with impressions of *Reticuloceras bilingue* showing that the succession ranges up into the R_{2b} subzone.

W. H. Morton.

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16th November, 1959.

THE BUGLE PIT, HARTWELL, NEAR AYLESBURY, BUCKS SP(42)793121

Sir,—The Bugle Quarry used to be the type section of the Portland and Purbeck Beds of the Aylesbury neighbourhood (Arkell, 1947, p. 126) and has been regularly visited by geologists since at least as early as 1854. It has long been out of work and has slowly become more and more obscured. Up to about three years ago, quite a good section of the Purbeck Beds could be seen. In 1958 an unexpected *coup-de-grâce* was given by tipping the spoil from extensive main road improvements over the face, which completely hid all

exposures with several feet of very mixed inert material, including much sand from the Wealden Beds of the adjacent Stone outlier.

In 1951 the Bugle Pit was notified to the Local Planning Authority by the Nature Conservancy as a *Site of Special Scientific Interest* under the terms of Section 23 of the National Parks and Access to the Countryside Act, 1949. In 1959 the Conservancy approached the land-owners, who kindly agreed to the re-opening of a permanent section, and in September some members of the Geology Department, Queen Mary College, excavated a section on the south-west face of the old quarry, now a thorn-tree covered bank, and it is intended to keep this exposure in good condition.

This section is some 9 feet wide, and exposes the uppermost 8 feet of Purbeck strata, down to within about a couple of feet of its base.

Intending visitors must obtain permission from the tenant farmer, Mr. L. J. Pitcher, Glebe Farm, Hartwell (SP(42)79451208) either at the time or preferably in advance. Apply at the farm buildings between the Bugle Horn Inn and the field gate immediately opposite the by-road signed to Hartwell House. Access is by this field gate and the old track leading into the quarry, and visitors are requested to keep to this track, and to be meticulous in avoiding damage to the agricultural land and crops which, with the old pit, remain in private ownership. They are also asked not to damage the barbed wire which has been strung round the new excavation to prevent cattle falling into it ; and to refrain from excessive hammering of this section.

In 1949, V. Wilson (p. 220) recorded algal limestone from the Purbeck Beds. This was not found in the recent excavation, but slabs of similar material occur round some filled-in trial holes in the field immediately to the west of the old quarry face.

J. F. Kirkaldy,
Hon. Sec. Geological Conservancy Council.

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Mile End Road,
London, E. 1.

14th December, 1959.

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REVIEWS

THE EARTH AND ITS GRAVITY FIELD. By W. A. HEISKANEN and F. A. VENING MEINESZ (McGraw Hill Series in the Geological Sciences). Published by McGraw Hill Book Co., Ltd., New York, 1958, 470 pp. Price 97*s*.

The launching of artificial Earth satellites has once again drawn attention to the importance of a knowledge of the Earth's gravitational field, for without such data the orbits cannot be accurately predicted. Oddly enough, this application is almost the only branch of the subject which the authors do not discuss. Otherwise the book is comprehensive, well-written, well illustrated, and has a good index.

After a short introduction and a general outline of the constitution of the Earth, the theory of the Gravity Field and the Figure of the Earth are presented. These chapters were mainly written by Vening Meinesz.

The technique of gravity measurements and the various methods of reducing them are then described. Considerable emphasis is laid on the value of isostatic reductions and two chapters are devoted to the use of gravity data for the solution of geodetic problems. Heiskanen was chiefly responsible for this part of the book.

In the last three chapters, Vening Meinesz describes some of the theories which have been put forward to account for deviation from isostatic equilibrium and touches on such problems as the origin of continents and oceans, geosynclinal belts, and polar wandering. This section is much more controversial than the other two, but will be read with great interest by geologists and geophysicists.

Most of the material in the book has already been published but the authors have done a valuable service by collecting it together and presenting it in the form of a logical and readable book. Lists of references are given at the end of each chapter and both name and subject indices are provided.

B. C. B.

ELEMENTS OF GEOLOGY. By JAMES H. ZUMBERGE. pp. 8 + 382 with 199 text-figs, 18 chapters, title page illustrations, and 12 tables. J. Wiley, New York, 1958. Chapman and Hall, London, Price 44*s*.

PRINCIPLES OF GEOLOGY (Second Edition) by JAMES GILLULY, A. C. WATERS, and A. O. WOODFORD. pp. ix + 534, with 326 text-figs. and tables. W. H. Freeman and Co. New York. Price 42*s*.

Both works are first courses evolved by widening the field covered in the standard college text in physical geology. *Elements* is expressly designed for a one semester course in both physical and historical geology for those whose studies are primarily directed elsewhere. Thus it may be considered as an introduction for general readers or in the United Kingdom as a school text. It is attractively produced but inevitably its stratigraphical half is directed to an American reader.

Principles is altogether more promising and possibly more disappointing. It has more than twice the substance and compared with most books of the same scope is very reasonably priced. It exceeds the content of the average physical geology in having a chapter on " Strata, fossils, and time " and other similar but lesser extensions. The chapter headings, layout, and diagrams give a good impression and suggest a serious competitor for an introductory undergraduate text, as indeed it is. However, like an excellent photograph which is out of focus, on closer examination the details are often blurred. Historical analysis is often superficial. Emphasis on general principles rather than on technical terminology is to be welcomed, but the inevitable jargon is not always used with a rigorous appreciation of the principles involved. A generalized approach is evident in controversial matters where extreme views are presented and rather than pointing the way to a critical analysis the reader is left with the idea that there is much to be said on both sides and with the

task of averaging out the differences. Thus *Principles* gives a good general impression of the present geological outlook but without the incisive argument which distinguished the first *Principles* as a classic of scientific as well as of historical method. To-day's *Principles* and *Elements* certainly impress with an array of knowledge, but perhaps inevitably lack the method of their illustrious predecessors.

W. B. H.

DARWIN'S BIOLOGICAL WORK: SOME ASPECTS RECONSIDERED. Edited by P. R. BELL. xiii + 343 pp., with 6 plates and numerous text-figures. Cambridge University Press, 1959. Price 40*s*.

This book, issued on the occasion of the centenary of the publication of *The Origin*, comprises six essays designed to illustrate something of the breadth of Darwin's interests and activities as a biologist: The movement of plants in response to light (P. R. Bell), Palaeontology and Evolution (J. Challinor), Natural Selection (J. B. S. Haldane), Developments in the study of animal communication (P. Marler), and Cross- and Self-fertilization in plants (H. L. K. Whitehouse). Of considerable interest to the general reader is the final essay, Buffon, Lamarck and Darwin: the originality of Darwin's Theory of Evolution, by J. S. Wilkie, which may appropriately be read in conjunction with H. G. Cannon's recently published *Lamarck and Modern Genetics* (Manchester University Press).

O. M. B. B.

PUBLICATIONS RECEIVED

G. F. WILLMOT, J. M. BIGGINS, and P. M. TILLOTT (Ed.). *York : a Survey, 1959*. 198 pp., published by the Local Executive Committee, British Association, York, 1959.

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—— *Annual Report of the Geological Survey Department for 1958*, by F. W. ROE. xii + 247 pp., with 44 figs., 32 tables and 50 plates. Kuching, Sarawak, 1959. Price 7*s*.

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